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GEOLOGICAL MAGAZINE

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Foraminiferal Faunas from the Lapithos Group of Cyprus [1]

By A. E. COCKBAIN

ABSTRACT

This paper reviews the stratigraphic micropalaeontology of the Lapithos Group of Cyprus, which is now known to be Upper Cretaceous to Lower Miocene in age. A three-fold subdivision, on the basis of planktonic foraminifera, into Lower (Upper Cretaceous) Middle (Danian to Eocene) and Upper (Oligocene to Lower Miocene) Lapithos Formations is suggested and a summary of the outcrops of each formation is given. The foraminiferal faunas of the Lapithos Group are discussed and it can be shown that a planktonic and a neritic faunal assemblage is present in each formation. The Terra Limestone, hitherto considered as part of the Dhali Group, shows faunal affinities with the Upper Lapithos neritic facies and is therefore placed in the Lapithos Group.

INTRODUCTION

THE Lapithos Group is the name given to those chalks, marls, and limestones occurring in Cyprus which are Upper Cretaceous to Lower Miocene in age. Macrofossils are extremely rare but the abundant microfauna—chiefly foraminifera—enables a three-fold subdivision of the Group to be made. The Lower Lapithos Formation includes Upper Cretaceous rocks and is limited to the Maestrichtian; the Middle Lapithos Formation ranges from the Danian and Palaeocene to the Eocene whilst the Upper Lapithos is Oligocene and Lower Miocene in age. On faunal grounds, two main facies may be recognized. The greater part of the chalks in the Lapithos Group have an assemblage of planktonic foraminifera but in some limestone beds a neritic fauna of larger foraminifera is present.

The Lapithos Group crops out in the Kyrenia Range, north of the Ovgos river, at Phano Hill and around the Troodos Mountains (see Browne and McGinty's map in Henson, Browne, and McGinty, 1949; also Text-fig. 1). The Kyrenia Range outcrops belong mainly to the Lower and Middle Lapithos Formations, with some Upper Lapithos occurring in the eastern part of the Range. Upper Lapithos is present in the Ovgos river section and probably only Middle Lapithos occurs at Phano Hill. All three formations are found on the flanks of the Troodos Mountains. The chalks on the northern and eastern sides of the Troodos Mountains have been examined in most detail and it was here that the three-fold subdivision was first established.

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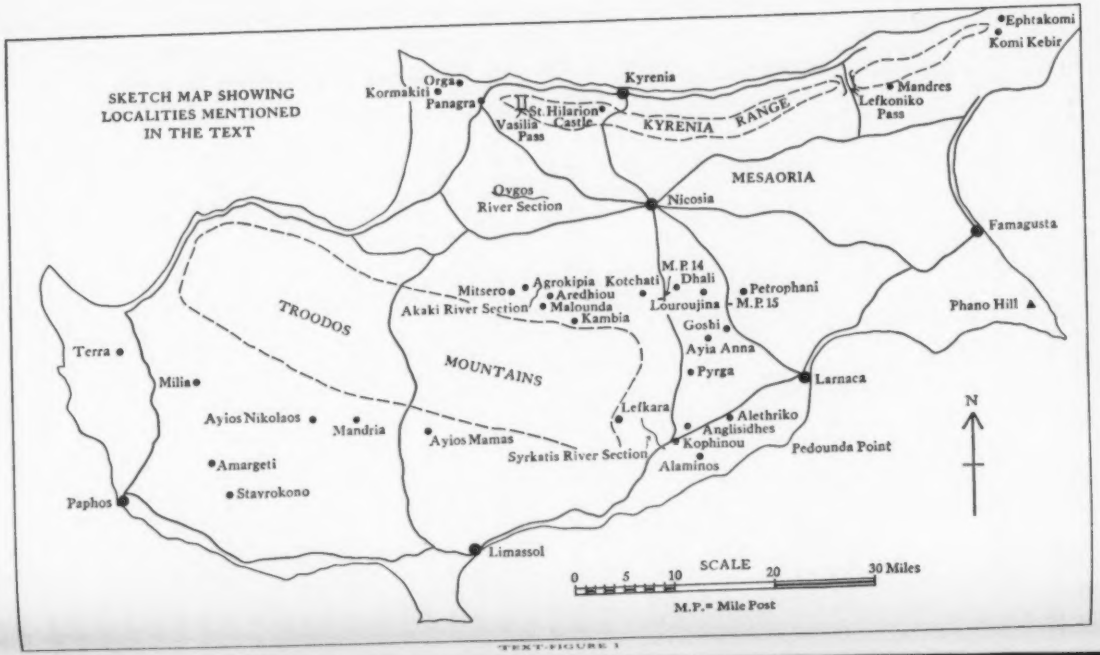

TEXT-FIGURE 1

The writer wishes to acknowledge the help given him in the preparation of this paper by his colleagues in the Geological Survey Department, Cyprus. In particular he would like to thank Messrs. Bagnall, Bear, Gass, and Moore for showing him several interesting outcrops and allowing him to draw freely on their field work. Without their assistance the present paper could not have been written.

Previous Work

A short synonymy of the Lapithos Group is given by Henson, Browne, and McGinty (1949) together with a brief review of the contributions of previous authors. Henson and his co-workers defined the Lapithos Group as comprising the predominantly calcareous beds of Upper Cretaceous to Oligocene age and were the first to publish an adequate account of the foraminiferal faunas. They abandoned the terms "Lapithos Formation" and "Ovgos Formation" previously used by Browne and McGinty (1946) for these rocks, considering the latter formation to be of Oligocene age and hence part of the Lapithos Group.

The term "Ovgos Formation" was resurrected by Wilson (in Ingham, 1958, p. 23) who said that it "... is apparently Oligocene in age and is believed to be the upper portion of the Lapithos Group...". This age was based on Henson, Browne, and McGinty's (1949) faunal list; as mentioned below, it is now thought that this fauna indicates a Miocene age. Since the formation has never been sufficiently defined, it is abandoned here, the beds being included in the Upper Lapithos Formation (see below).

Later work has shown that the top of the Lapithos Group is Lower Miocene in age. Dr. C. G. Adams of the British Museum (Natural History), in a report (in Ingham, 1958), pointed out that the presence of *Orbulina universa* in the Ovgos rocks indicates a Miocene age and he recorded the same fossil from some localities around the Troodos Mountains. Recent work (Cockbain in Ingham, 1959; in Bear, 1959; in Gass, in press; etc.) has extended the area covered by Lower Miocene Lapithos.

Classification

In the absence of macrofossils, a palaeontological subdivision of the Lapithos Group must be based on the microfauna. Although both foraminifera and ostracods occur, the former are far more abundant and the subsequent discussion will deal exclusively with them. The planktonic foraminifera have been found especially useful since the sequence of these fossils in Upper Cretaceous to Lower Miocene times is well established in many parts of the world.

Various subdivisions of the Lapithos Group, based on lithology,

were made by the officers of the Cyprus Geological Survey (see Ingham, 1956, 1957, 1958, 1959) but in the absence of palaeontological control these groupings were of only local significance. Within the Dhali Area, Gass (in Ingham, 1957; Gass, in press) was able to map several units in the Lapithos chalks on lithological and structural grounds. Subsequent examination of the foraminiferal faunas confirmed the major subdivisions identified in the field and has enabled the island-wide recognition of three formations:

(1) Lower Lapithos Formation. This is of Upper Cretaceous (Maestrichtian) age and is mainly a chalk or marly chalk sequence. The planktonic foraminifera include *Globotruncana* and *Heterohelix*. A neritic fauna of orbitoids and calcarinids is also found in some areas.

(2) Middle Lapithos Formation. Danian, Palaeocene, and Eocene strata are grouped together in this formation. Following Reiss (1955), Troelsen (1957), and Loeblich and Tappan (1957) the Danian is considered Tertiary and not Cretaceous. Chalks with chert make up the bulk of this formation. The lower part of the Middle Lapithos contains a *Truncorotalia* fauna; in higher beds *Hantkenina* and *Globigerapsis* are found although they are by no means abundant. Larger foraminifera such as alveolinids, discocyclinids and nummulites are common in certain limestone beds.

(3) Upper Lapithos Formation. The Oligocene and Lower Miocene are both included here. Lithologically the succession is uniform, consisting of bedded chalks with occasional marls. The formation contains a good benthonic fauna with *Cassidulina*, *Cibicides*, *Planulina*, and *Parrella* as characteristic genera. This is in direct contrast to the previous formations in which benthonic genera are not common. Planktonic species include *Globigerina rohri*, *G. venezuelana*, *Globoquadrina altispira*, and *Orbulina universa*. Chattian–Aquitanian faunas of orbitoids and other larger foraminifera are present in the Upper Lapithos Formation.

This three-fold subdivision closely corresponds to the tentative differentiation of the Lapithos Group suggested by Henson, Browne, and McGinty (1949). These authors thought that at least two formations could be recognized on the presence of a break between the Maestrichtian and Lutetian (Middle Eocene), and further put forward the possibility that certain sandy Upper Eocene and Oligocene beds in the eastern part of the Kyrenia Range might be mapped separately. The unconformity between the Maestrichtian and Lutetian is the break which frequently separates the Lower from the Middle Lapithos Formation. The sandy Upper Eocene and Oligocene Lapithos beds have not been examined but it seems that they correspond approximately to the Upper Lapithos, as here defined. The Upper Eocene is

placed in the Middle Lapithos since it lacks the benthonic fauna characteristic of the Upper Lapithos Formation.

In the following paragraphs, the foraminiferal faunas from the three formations of the Lapithos Group will be discussed in some detail. Two tables summarizing the distribution of the foraminifera in these formations are given at the end of this paper.

Lower Lapithos Formation

The oldest Lapithos rocks are of Maestrichtian age and are placed in the Lower Lapithos Formation. Outcrops of this formation are confined to the Kyrenia Range and certain localities around the Troodos Mountains.

In the Kyrenia Range, the contact between the Lapithos Group and older rocks is nearly always faulted (see, for example, Moore in Ingham, 1959). However, in at least one exposure, near St. Hilarion Castle, a conglomerate occurs between Lower Lapithos chalks and Hilarion Limestone and this may indicate an unconformity. In addition to this locality, Maestrichtian Lapithos has been recorded from Panagra, Vasilia Pass, near Orga, north of Mandres, Lefkoniko Pass, and around Komi Kebir.

The planktonic faunal assemblage occurs in red, green, or whitish, frequently schistose, chalks. Because of the indurated nature of the chalk the foraminifera are difficult to extract and the Maestrichtian fauna is less well known in the Range as compared with the Troodos outcrops. One of the best faunas is recorded from near Orga and consists of *Cibicides* sp., *Globotruncana stuarti*, small globigerinids, *Heterohelix* sp., and *Racemiguembelina fructicosa* (Cockbain in Moore, MS.). From chalks cropping out in the Lefkoniko Pass, Dr. Z. Reiss of the Israel Geological Survey (private report) has identified *Globotruncana contusa*, *G. gansseri*, *G. stuarti*, and *Pseudotextularia* spp. Elsewhere the fauna has been examined only in thin sections yielding *Globotruncana* spp., globigerinids, and *Heterohelix* spp.

Larger foraminifera have been recorded, mainly from the eastern end of the Kyrenia Range, by Renz and Reichel (1946) and Henson, Browne, and McGinty (1949). The faunas include *Orbitoides media*, *Siderolites calcitrapoides*, *Simplorbites* sp. and *Lepidorbitoides* spp. These foraminifera are found in limestones which are often very difficult to separate from the Hilarion Limestone. This difficulty is enhanced by the uncertainty as to the upper age limit of the Hilarion Limestone. Renz (1929) records Maestrichtian foraminifera from the Trypanian (that is the Hilarion Limestone) and it is possible that both the upper portion of the Hilarion Limestone and the Lower Lapithos Formation are of this age, in which case some criterion, other than age, for distinguishing them must be found.

Around the Troodos Mountains the Lapithos Group may rest unconformably on pillow lavas or the Mamonia Formation, or apparently conformably on the Perapedhi Formation. The Lower Lapithos passes down into the Perapedhi Formation with no visible break and it is difficult to find the junction between them. For this reason, some workers (see Gass, in press) have suggested that the beds of the Perapedhi Formation could be considered as part of the Lower Lapithos Formation. This possibility awaits palaeontological confirmation, since the radiolaria on which the supposedly Jurassic age of the Perapedhi is based are unreliable as age indicators.

Outcrops of Lower Lapithos are found where erosion has cut through the overlying transgressive Middle Lapithos Formation. Those known at present occur at Milia, near Ayios Nikolaos, in the Syrkatis river section and near Anglisidhes, Pyrga, and Lymbia. The chalk at these localities is often pink in colour, occasionally marly, and may contain chert.

The white chalk at Milia in the Paphos District (Cockbain in Ingham, 1959) has yielded a good planktonic fauna with *Globotruncana gagnebini*, *Heterohelix globulosa*, *Planoglobulina* sp., *Pseudotextularia elegans*, *Racemiguembelina fructicosa* and *Rugoglobigerina rugosa*. Also on the south side of the Troodos Mountains, Henson, Browne, and McGinty (1949) have recorded an Upper Cretaceous foraminiferal fauna from the Ayios Nikolaos–Mandria road section. The fossils probably came from the Lower Lapithos marl division of Wilson (Wilson and Ingham, 1959) and include the Maestrichtian species *Globotruncana cretacea*, *Pseudotextularia elegans* (as *Gümbelina elegans*) and *Racemiguebelina fructicosa* (as *Pseudotextularia varians*) (see Gallitelli, 1957, for a revision of the taxonomy of the Heterohelicidae). Arenaceous genera, for example *Dorothia*, *Gaudryina*, and *Spiroplectammina*, also occur at this locality.

The only outcrop of the Lower Lapithos Formation known on the north side of the Mountains is at Lymbia. The formation here consists of marls and chalks and the fauna includes *Globotruncana cretacea*, *G. stuarti*, *Pseudotextularia elegans*, *Dorothia* sp., and *Textularia* sp. (Cockbain in Gass, in press). Adams (in Ingham, 1958) also gives a list of foraminifera from this locality. On the eastern flank of the Troodos Mountains, the Lower Lapithos Formation again crops out near Pyrga and further south at Anglisidhes where chalks with tuff bands have yielded *Globotruncana* cf. *arca*, *G. gansseri*, *G. stuarti*, and the large foraminifer *Orbitoides* sp. (Cockbain in Bagnall, in press). *Globotruncana* has been recorded from the base of the Syrkatis river section (Henson, Browne, and McGinty, 1949).

Although the planktonic foraminiferal assemblage is well represented in the Lower Lapithos Formation around the Troodos Mountains,

evidence of the neritic assemblage is found only at Anglisidhes. There seems to be a slight variation within the planktonic assemblage dependant on the amount of marl present in the chalks. In the Ayios Nikolaos and Lymbia exposures, where marly chalks occur, the fauna includes arenaceous foraminifera. The predominantly chalk localities, such as Milia and Anglisidhes, lack these forms.

Middle Lapithos Formation

The boundary between the Lower and Middle Lapithos Formations is taken at the world-wide break between the Maestrichtian and Danian (see Reiss, 1955). In many places Danian and Palaeocene strata are missing, the Eocene resting directly on the Maestrichtian with apparent conformity (Henson, Browne, and McGinty, 1949; Gass, in press; Bagnall, in press). Furthermore, the formation is transgressive and Danian, Palaeocene, or Eocene beds may overlie the pillow lavas along the north side of the Troodos Mountains (Bear, 1960).

The Middle Lapithos Formation is mainly a chalk with chert succession. Although chert is sometimes found in the Lower Lapithos, it is essentially characteristic of the Middle Lapithos. Certain lithological variations are present, for example in the Dhali Area (Gass, in press) where massive chalks occur in the upper part of this formation.

Middle Lapithos beds crop out in the Kyrenia Range, around the Troodos Mountains and at Phano Hill. Of these outcrops, those around the Troodos Mountains have been studied in most detail.

Along the northern flank of the Troodos Mountains the formation extends from near Mitsero in the west to Louroujina in the east (Bear, 1960; Gass, in press). On the eastern side of the Massif the formation has been mapped between Louroujina and Kophinou (Gass, in press; Bagnall, in press) and on the south side round to Lefkara (Bagnall, in press). Further west the formation has not yet been mapped, but rocks of this age are known from the Paphos District (Cockbain in Ingham, 1959) and the Ayios Nikolaos–Mandria road section (Henson, Browne, and McGinty, 1949) where they probably correspond to Wilson's "Lapithos Chalk" (Wilson and Ingham, 1959) which extends from Ayios Nikolaos eastwards to Ayios Mamas.

In the Kyrenia Range, the Middle Lapithos Formation has been identified in the Kormakiti Area (Moore, MS.) and at several localities along the Range where Renz and Reichel (1946) found Palaeocene and Eocene strata; the faunal lists of the latter authors suggest that at least some of their Palaeocene outcrops are of Eocene age.

The Phano Hill outcrop has been described by Browne in a private report; a passing reference is made to it by Henson, Browne, and McGinty (1949).

The planktonic foraminifera of the Middle Lapithos Formation may

be divided into four assemblages characterized respectively by *Globorotalia pseudobulloides*, *Truncorotalia*, *Hantkenina*, and *Globigerapsis*. These assemblages enable Danian, Palaeocene, Lower and Middle Eocene, and Middle and Upper Eocene strata to be recognized.

(1) *Globorotalia pseudobulloides* assemblage. This assemblage is recorded from near Aredhiou where it consists of poorly preserved specimens of *Globorotalia pseudobulloides*, *Cibicides* sp., *Gyroidina* sp., and *Marssonella* sp. (Cockbain in Bear, 1960). *G. pseudobulloides* occurs in the Danian and lower Palaeocene (Loeblich and Tappan, 1957); in the absence of *Truncorotalia*, which first appears in the Palaeocene, the Aredhiou outcrop is considered to be of Danian age.

G. pseudobulloides, associated with *Globigerina triloculinoides* also occurs in the " Upper Cretaceous " fauna recorded from the Ayios Nikolaos–Mandria section by Henson, Browne, and McGinty (1949). The rest of the " Upper Cretaceous " fauna from this section is here considered to be of Maestrichtian age (see above) but these two species are taken to indicate the presence of Danian strata.

(2) *Truncorotalia* assemblage. This is the most widespread Middle Lapithos faunal assemblage. It has been found at many places around the Troodos Mountains and also in the Kyrenia Range. The assemblage is Palaeocene to Middle Eocene in age.

Palaeocene Lapithos outcrops are few in number. On the north side of the Troodos Massif the only known occurrence is near Mitsero where *Truncorotalia velascoensis* is found together with *Globigerina triloculinoides* and *Cibicides pseudoungerianus* (Cockbain in Bear, 1960). The Palaeocene may be present south of the Massif in the Ayios Nikolaos–Mandria section (Henson, Browne, and McGinty, 1949) where *Globigerina* spp. and *Globorotalia* spp. are recorded from beds of " Danian or Palaeocene to Middle Eocene " age.

In the eastern end of the Kyrenia Range, the *Truncorotalia* assemblage has been identified by Renz and Reichel (1946) who gave it a Palaeocene age. The species found include *Truncorotalia aragonensis*, *Globorotalia spinulosa*, and *G.* cf. *crassula*, which most workers would now consider to be of Eocene age.

Truncorotaliae of Lower and Middle Eocene age are found at many localities around the Troodos Mountains. The best known faunas of this age come from the north side of the Mountains. At Agrokipia, *Truncorotalia aragonensis*, *T.* cf. *rex*, and *T. wilcoxensis* are found associated with *Globigerina soldadoensis*, *Robulus* sp., *Cibicides pseudoungerianus*, and *Eponides umbonatus* var. *stellatus* (Cockbain in Bear, 1960). Further east, on the Nicosia–Limassol road (milepost 14) *T. aragonensis* and *G. soldadoensis* have been recorded. Henson, Browne, and McGinty (1949) mention the presence of Eocene Lapithos near Dhali. In the area around Dhali several exposures have yielded

good faunas of the *Truncorotalia* assemblage. The foraminifera include *T. aragonensis*, *T. aragonensis* var. *caucasica*, *T. quetra* group, and *T. wilcoxensis* (Cockbain in Gass, in press).

On the eastern side of the Mountains, *Truncorotalia* faunas have been recorded from north-east of Anglisidhes and near Alethriko. At the latter locality, Dr. Z. Reiss has identified *Truncorotalia crassata* and *T. wilcoxensis* (listed in Ingham, 1958). North of Kophinou, on the Nicosia–Limassol road, the *Truncorotalia* fauna is found in association with the *Hantkenina* assemblage.

Along the southern flank of the Troodos Mountains the fauna is found near Lefkara, where *T. aragonensis* and *G.* cf. *soldadoensis* have been identified and it possibly occurs in the Ayios Nikolaos–Mandria section (Henson, Browne, and McGinty, 1949). In the Paphos District *Truncorotalia* and *Hantkenina* have been found together (Cockbain in Ingham, 1959).

At the western end of the Kyrenia Range an Eocene *Truncorotalia* assemblage is present north of Panagra. The fauna here includes *T. aragonensis*, *T. aequa*, *T. wilcoxensis*, and *G.* cf. *soldadoensis* (Cockbain in Moore, MS.).

(3) *Hantkenina* assemblage. The *Hantkenina* assemblage, of Middle and Upper Eocene age, has a restricted occurrence, having been found only in the Paphos District and in the neighbourhood of Kophinou. Near Amargeti, in the Paphos District, *Hantkenina dumblei* is associated with *Truncorotalia* cf. *bullbrooki* and *Turborotalia* cf. *centralis* (Cockbain in Ingham, 1959). North of Kophinou, two *Hantkenina* faunas are found. The lower one contains *H. mexicana* together with *Truncorotalia aragonensis*, *T.* cf. *quetra*, *Globigerina* cf. *soldadoensis*, and *G. primitiva* and is of Middle Eocene age. Above it, the Upper Eocene *H. alabamensis*, *Cribrohantkenina bermudezi*?, and *Turborotalia centralis* have been identified. *Hantkenina* is also recorded from the nearby Syrkatis river section (Henson, Browne, and McGinty, 1949).

(4) *Globigerapsis* assemblage. This assemblage also has a restricted occurrence, being confined to the massive chalks of the upper portion of the Middle Lapithos Formation in the Dhali Area. The fauna consists of *Globigerapsis index*, *G. kugleri*, *Globigerina* cf. *soldadoensis*, *Globorotalia* cf. *aspensis*, *Cibicides grimsdalei*, and *C.* cf. *mexicanus*; it is considered to be Middle and Upper Eocene in age (Cockbain in Gass, in press). At present the relationship of this fauna to the *Hantkenina* assemblage is not known since the two have not been found in association.

Although not belonging to the *Globigerapsis* assemblage, two other Middle and Upper Eocene faunas may be mentioned here. Both were recorded by Henson, Browne, and McGinty (1949). North-east of

Kophinou these authors found *Porticulasphaera mexicana* and *Globigerina cryptomphala*; these foraminifera probably came from the same beds as the Kophinou *Hantkenina* faunas referred to above. Certain chalks in the Ayios Nikolaos–Mandria section have yielded *G. cryptomphala*, *Globigerinoides conglobatus*, and several species of benthonic foraminifera. It is possible that these fossils were collected from the "Lapithos Chalk" (see Wilson in Wilson and Ingham, 1959).

Benthonic foraminifera are not common in the Middle Lapithos Formation. Besides the nearly ubiquitous *Cibicides pseudoungerianus* and *Eponides umbonatus* var. *stellatus*, the most commonly occurring specimens are globose forms of *Cibicides* such as *C. grimsdalei* and *C. mexicanus*. Other species include *Cibicides* cf. *robertsonianus*, *Eponides* cf. *trumpyi*, *Pullenia bulloides*, and *Robulus* sp.

Assemblages of large foraminifera of Middle Lapithos age are present in the Kyrenia Range, at Phano Hill and around the Troodos Mountains between Ayia Anna and Anglisidhes. Renz and Reichel (1946) have recorded *Neoalveolina*, *Miscellanea*, and *Discocyclina* from several localities in the Kyrenia Range. At Ephtakomi, Henson, Browne, and McGinty (1949) found *Nummulites gizehensis*, *N. atacicus*, *Assilina* sp., and *Discocyclina* spp. More recently, cherty limestones with nummulites and discocyclinids have been found near St. Hilarion Castle and Orga. According to Henson, Browne, and McGinty (1949) the Lapithos at Phano Hill contains nummulites.

Nummulites, *Discocyclina*, *Asterocyclina*, and *Alveolina* are present in tuff bands within the Middle Lapithos between Ayia Anna and Anglisidhes (Cockbain in Gass, in press; in Bagnall, in press). The surrounding chalks contain a *Truncorotalia* fauna of Eocene age. This fauna of larger foraminifera may have been previously recorded by Henson, Browne, and McGinty (1949) who mention the presence of *Nummulites*, *Assilina*, *Discocyclina*, and *Linderina* in the "... Eocene limestones and cherts at several places ..." presumably along the north side of the Troodos Mountains.

Upper Lapithos Formation

The lower boundary of the Upper Lapithos Formation is taken at the unconformity which separates it from the Middle Lapithos in the Dhali Area (Gass, in press) and also probably further west (Bear, 1960). In the Dhali Area the Upper Lapithos may rest on beds carrying either the *Globigerapsis* or the *Truncorotalia* assemblage and in the west it oversteps onto the pillow lavas (Gass, in press). The most noteworthy palaeontological feature of the Middle-Upper Lapithos junction is the increase in benthonic foraminifera in the Upper Lapithos. This increase marks the first appearance of a good benthonic foraminiferal

fauna in Cyprus; it continues, with only slight modifications, into the Dhali Group.

Lithologically the Upper Lapithos Formation is fairly uniform, consisting of bedded chalks and marls with no chert. An areally extensive slump horizon occurs in the upper part of the formation (Gass, in press; Bagnall, in press).

The upper limit of this formation is not always easy to draw either in the field or on faunal grounds. Stratigraphically the formation may pass conformably up into the Kythrea Formation, as in the Ovgos river section or it may be unconformably overlain by the Pakhna Formation, as in many places around the Troodos Mountains. The Upper Lapithos, Kythrea, and Pakhna Formations all contain a similar fauna. In particular the Upper Lapithos planktonic fauna continues almost unchanged into the Dhali Group. The main differences are that the Dhali Group formations have a more varied buliminid and lagenid fauna with *Bolivina arta*, *B. catanensis*, *Hopkinsina bononiensis* *Lagena* sp., and *Lagenonodosaria scalaris*, and yield *Amphistegina* which is not known from the Lapithos. In addition, *Sphaeroidinella* is quite common in the Pakhna and has also been found in the Kythrea but never in the Lapithos.

Outcrops of the Upper Lapithos Formation are present in the Kyrenia Range, the Ovgos river section and around the Troodos Mountains. Henson, Browne, and McGinty (1949) record sandy Oligocene beds from the eastern part of the Kyrenia Range and these represent an arenaceous facies of this formation. The fauna from the Ovgos river section (Henson, Browne, and McGinty, 1949) indicates that the cleaved chalks and marls cropping out here belong to the Upper Lapithos Formation.

The formation has been mapped along the north and east sides of the Troodos Mountains from Mitsero to Goshi and south as far as Alaminos (Bear, 1960; Gass, in press; Bagnall, in press). On the southern flank of the Mountains, Upper Lapithos is present south of Lefkara (Bagnall, in press). Further west it is probably represented, at least in part, by the Upper Lapithos marls of Wilson (Wilson and Ingham, 1959) between Ayios Nikolaos and Ayios Mamas. The formation has also been recognized in the Paphos District (Cockbain in Ingham, 1959).

Faunally, the Upper Lapithos may be subdivided into two. The lower beds contain *Globigerina rohri* and *G. venezuelana* and are Oligocene in age; the upper portion is of Lower Miocene age (Aquitanian and Burdigalian) and is characterized by *Globoquadrina* and *Orbulina*.

(1) Oligocene beds. Good exposures of these beds occur near Mitsero. The foraminiferal fauna includes *Globigerina rohri*, *G.* cf.

trilocularis, *G. venezuelana*, ?*Globorotaloides suteri*, *Bolivinopsis clotho*, *Cassidulina subglobosa*, *Parrella mexicana*, *Planulina renzi*, and *Siphonodosaria* sp. (Cockbain in Bear, 1960). An exposure in the Akaki river further east yields *Globigerina ciperoensis*?, *G. rohri*, *G. venezuelana*, and *Globorotalia kugleri* (Cockbain in Bear, 1960). In the Dhali Area, Oligocene rocks are well exposed on the flanks of the Petrophani igneous inlier (Cockbain in Gass, in press). The fauna from this locality includes *Catapsydrax dissimilis*, *Globigerina* cf. *ciperoensis*, *G. rohri*, *G. venezuelana*, *Cibicides cicatricosus* var. *maioricensis*, and *Siphonina reticulata*. Exposures south of Lefkara have yielded a similar fauna. Along most of the southern flank of the Troodos Mountains the Oligocene faunas are, as yet, poorly known.

Adams (in Ingham, 1958) has identified a fauna of "Eocene-Oligocene" age from Lapithos chalks collected on the eastern side of the Mountains. Although there are no diagnostic species in his list, the general aspect of the fauna warrants its placing in the Oligocene part of the Upper Lapithos Formation.

(2) Lower Miocene beds. Lapithos chalks of Lower Miocene age were first recognized from Alaminos by Adams (in Ingham, 1958) who also pointed out that the chalks in the Ovgos river section were of the same age. Subsequently Lower Miocene strata have been found at numerous localities around the Troodos Mountains. Two faunas may be recognized, characterized by the presence or absence of *Orbulina*. The earlier fauna, without *Orbulina*, is found at Stavrokono in the Paphos District (Cockbain in Ingham, 1959) and below the Pakhna Formation at the Petrophani igneous inlier (Cockbain in Gass, in press). The Stavrokono exposure has yielded *Globoquadrina altispira*, *G. dehiscens*, and *Globigerinoides bispherica*. At Petrophani, *Globigerinoides bispherica* is associated with a typical Upper Lapithos benthonic fauna of *Anomalina* sp., *Cassidulina subglobosa*, *Parrella mexicana*, and *Planulina* cf. *renzi*. The presence of *G. bispherica* suggests that these exposures are at an horizon just below the "*Orbulina* surface" and are Aquitanian in age (Blow, 1956; 1957).

Higher strata are found, for example, near Kambia (Cockbain in Bear, 1960) where *Globigerinoides trilobus*, species of the *Globorotalia fohsi*, *G. mayeri*, and *G. menardi* groups and *Orbulina universa* occur, together with several benthonic species. Further east, at Kotchati (Cockbain in Ingham, 1959) the fauna includes *Biorbulina bilobata*, *Globoquadrina altispira*, *G. dehiscens*, *Orbulina suturalis*, and *O. universa*. From an exposure near Alaminos, Adams (in Ingham, 1958) identified *Anomalina* sp., *Eponides* sp., *Globigerina* cf. *dubia*, ?*Globigerinoides* cf. *trilobus*, *Globorotalia scitula*, and *Orbulina universa*. Other outcrops containing the *Orbulina* fauna occur near Mitsero, Dhali, Goshi, and Pedounda Point.

The foraminiferal fauna from the Ovgos river section is a typical Upper Lapithos fauna. The species include *Globoquadrina altispira*, *G. dehiscens*, *Orbulina universa*, and numerous benthonic foraminifera (see Henson, Browne, and McGinty, 1949). Browne (private report) has suggested a Burdigalian age for the succession and the fauna supports this suggestion.

Towards the top of the Upper Lapithos Formation there is a slump horizon. The fauna from an exposure of this horizon on the Nicosia–Larnaca road (milepost 15) consists of *Globoquadrina altispira*, *Orbulina universa*, *Anomalina* sp., *Cassidulina subglobosa*, *Cibicides pseudoungerianus*, *Eponides umbonatus* var. *stellatus*, and *Siphonina reticulata*. The age of these slumped beds and the other strata bearing the *Orbulina* fauna is upper Aquitanian and Burdigalian.

An examination of the fauna of the Upper Lapithos Formation indicates that the uppermost beds are not the same age everywhere. For example, the slump horizon cropping out on the Nicosia–Larnaca road contains *Orbulina*, whereas in the Petrophani igneous inlier the highest Lapithos beds carry *Globigerinoides bispherica* but not *Orbulina* (Cockbain in Gass, in press); near Aredhiou, Mitsero, and in the Akaki river the uppermost Lapithos is Oligocene but at Malounda and Agrokipia Lower Miocene strata are present (Cockbain in Bear, 1960). From these observations it follows that there must have been a certain amount of uplift and erosion prior to the deposition of the Pakhna sediments.

Neritic faunas of Oligocene and Lower Miocene age are known from two areas in Cyprus. North of Komi Kebir *Nephrolepidina sumatrensis*, *Eulepidina laevis*, *E. dilatata*, *Miogypsinoides complanatus*, *Spiroclypeus* sp., and *Heterostegina* sp. were discovered by Renz and Reichel (1946). They considered this fauna to be Oligocene in age and to occur in the Kythrea Formation. Henson, Browne, and McGinty (1949), who later re-examined the locality, gave a Chattian–Aquitanian age to the fauna and suggested that it came from the upper part of the Lapithos Group.

In the southern part of the island a reef deposit of Chattian–Aquitanian age has been mapped by Henson, Browne, and McGinty (1949) who called it the Terra Limestone and placed it in the Dhali Group. The fauna includes *Nephrolepidina marginata* = *tournoueri*, *Miogypsinoides complanatus*, *Heterostegina assilinoides*, and *Spiroclypeus* sp. and bears a close resemblance to the Komi Kebir fauna detailed above. From what is now known of the upper age limit of the Lapithos Group it seems logical to remove the Terra Limestone from the Dhali Group and consider it as a reef facies of the Upper Lapithos Formation.

TABLE 1.—VERTICAL DISTRIBUTION OF FORAMINIFERA OF THE NERITIC FAUNAL ASSEMBLAGE IN THE LAPITHOS GROUP OF CYPRUS

		Maestrichtian		Eocene		Chattian–Aquitanian	
		Kyrenia Range	Troodos	Kyrenia Range	Troodos	Kyrenia Range	Terra Limestone
Upper Lapithos Formation	*Amphistegina* sp.					X	X
	Eorupertia sp.						X
	Eulepidina dilatata (Michelotti)					X	
	E. laevis Douvillé					X	
	Heterostegina assilinoides Blanckenhorn emend. Henson						X
	Miogypsinoides complanatus (Schlumberger)					X	X
	Nephrolepidina sumatrensis Brady					X	
	N. marginata-tournoueri Lemoine & Douvillé						X
	Operculina sp.						X
	Operculinoides sp.						X
	Rotalia viennoti Greig						X
	Spiroclypeus sp.					X	X
Middle Lapithos Formation	*Alveolina* sp.				X		
	Assilina sp.			X	X		
	Asterocyclina sp.				X		
	Discocyclina sp.			X	X		
	Linderina sp.				X		
	Neoalveolina sp.			X			
	Nummulites spp.			X	X		
	N. atacicus Leymerie			X			
	N. gizehensis Forskål			X			
Lower Lapithos Formation	*Lepidorbitoides* spp.	X					
	Orbitoides sp.	X	X				
	O. media d'Archiac	X					
	Siderolites calcitrapoides Lamarck	X					
	Simplorbites sp.	X					

Compiled from data given by Renz and Reichel (1946), Henson, Browne, and McGinty (1949) and Cockbain in Ingham, (1959).

TABLE 2.—VERTICAL DISTRIBUTION OF FORAMINIFERA OF THE PLANKTONIC FAUNAL ASSEMBLAGE IN THE LAPITHOS GROUP OF CYPRUS

		Lower Lapithos Formation	Middle Lapithos Formation				Upper Lapithos Formation	
		Maestrichtian	Danian	Palaeocene	Lower & Middle Eocene	Middle & Upper Eocene	Oligocene	Lower Miocene
Upper Lapithos Formation	*Anomalina* sp.							X
	Robulus rotulatus (Lamarck)							X
	Uvigerina sp.							X
	Biorbulina bilobata (d'Orbigny)							X
	Globigerinoides bispherica Todd							X
	G. sacculifer (Brady)							X
	Globoquadrina altispira (Cushman & Jarvis)							X
	G. dehiscens Chapman, Parr, & Collins							X
	Globorotalia fohsi Cushman & Ellisor group							X
	G. menardii (d'Orbigny) group							X
	Orbulina suturalis Bronnimann							X
	O. universa d'Orbigny							X
	Cassidulina subglobosa Brady						X	X
	Cibicides boueanus (d'Orbigny)						X	X
	C. cicatricosus (Schwager) var. *maioricencis* Colom						X	X
	C. floridanus (Cushman) var. *miocenicus* Colom						X	X
	Gyroidina laevigata d'Orbigny						X	X
	Parrella mexicana (Cole)						X	X
	Planulina renzi Cushman & Stainforth						X	X
	Robulus curviseptus (Seguenza)						X	X
	Siphonina reticulata (Czjzek)						X	X
	Siphonodosaria sp.						X	X
	Vulvulina pennatula (Batsch)						X	X
	Globigerinoides trilobus (Reuss)						X	X
	Globorotalia mayeri Cushman & Ellisor group						X	X
	Bolivinopsis clotho (Gryzbowski)						X	
	Ehrenbergina sp.						X	
	Pleurostomella alternans (Schwager)						X	
	Pullenia bulloides (d'Orbigny)						X	
	Catapsydrax dissimilis (Cushman & Bermudez)						X	
	Globigerina cf. *ciperoensis* Bolli						X	
	G. rohri Bolli						X	
	G. venezuelana Hedberg						X	
	Globorotalia kugleri Bolli						X	
	G. opima Bolli						X	

Middle Lapithos Formation			Maestrichtian	Danian	Palaeocene	Lower & Middle Eocene	Middle & Upper Eocene	Oligocene	Lower Miocene
			Lower Lapithos Formation		Middle Lapithos Formation			Upper Lapithos Formation	
	Cibicides mexicanus Nuttall	*Globigerapsis* Assemblage					X	X	
	Eponides trumpyi Nuttall						X		
	Globigerapsis index (Finlay)						X		
	G. kugleri Bolli, Loeblich & Tappan						X		
	Hantkenina alabamensis Cushman	*Hantkenina* Assemblage					X		
	H. dumblei Weinzierl & Applin					X			
	H. mexicana Cushman					X			
	Eponides umbonatus (Reuss) var. *stellatus* (Silvestri)	*Truncorotalia* Assemblage				X	X	X	X
	Globigerina soldadoensis Bronnimann					X			
	Truncorotalia aragonensis (Nuttall)					X			
	T. quetra (Bolli)				X				
	T. rex (Martin)				X				
	T. wilcoxensis (Cushman & Ponton)					X			
	Globigerina triloculinoides Plummer				X				
	Truncorotalia velascoensis (Cushman)				X				
	Cibicides pseudoungerianus (Cushman)				X	X	X	X	X
	Cibicides sp.	*G. pseudobulloides* Assemblage							
	Gyroidina sp.			X					
	Marssonella sp.			X					
	Globorotalia pseudobulloides (Plummer)			X					

		Lower Lapithos Formation		Middle Lapithos Formation			Upper Lapithos Formation	
		Maestrichtian	Danian	Palaeocene	Lower & Middle Eocene	Middle & Upper Eocene	Oligocene	Lower Miocene
Lower Lapithos Formation	*Dorothia* sp.	X						
	Textularia sp.	X						
	Vulvulina sp.	X						
	Globotruncana cretacea Cushman	X						
	G. gagnebini Tilev	X						
	G. stuarti (de Lapparent)	X						
	Heterohelix globulosa (Ehrenberg)	X						
	Planoglobulina sp.	X						
	Pseudotextularia elegans (Rzehak)	X						
	Racemiguembelina fructicosa (Egger)	X						
	Rugoglobigerina rugosa (Plummer)	X						

Compiled from data given by Cockbain in Ingham (1959), in Bear (1960), and in Gass (in press).

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A "Vein" of Trondhjemite aplite : Granitization of Quartzite forming the sole of a Fold-thrust, Malin Head, Co. Donegal

By Doris L. Reynolds

(PLATE IX)

Abstract

A narrow layer of quartzite, emplaced within epidiorite by fold-thrusting, is transformed to albite-trondhjemite aplite and contiguous zones of the epidiorite are enriched in biotite. By reference to six chemical analyses it is found that, with but small additions of potash these transformations can be accounted for by the combined processes of (*a*) mechanical mixing of the initial quartzite and epidiorite, and (*b*) metamorphic inter-diffusion between this mixture and the adjoining epidiorite.

I Introduction

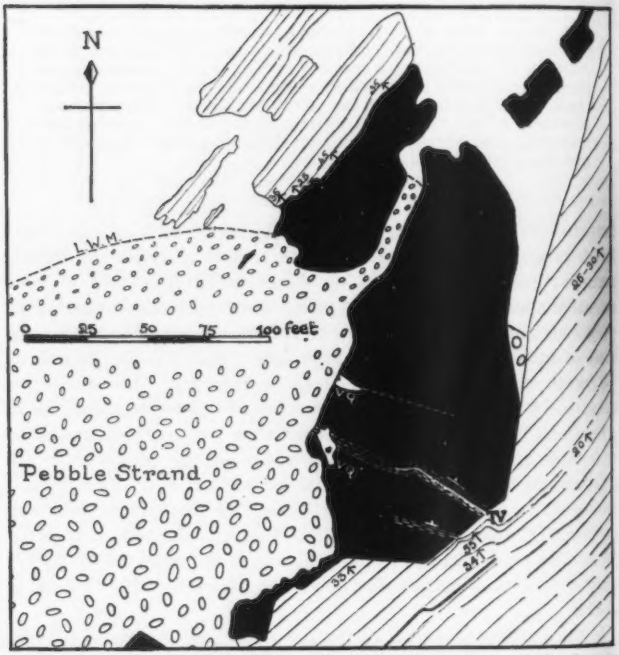
Interlayered with quartzite, the predominant rock of the Malin Head promontory, are relatively rare sheets of epidiorite. These occur at particular horizons as disconnected fragments of once extensive sheets that have been pulled apart by tectonic movements (Reynolds and Holmes, 1954). The present investigation relates to a sheet of epidiorite outcropping at the eastern end of Pebble Strand, on the north coast of the Malin Head promontory. The epidiorite, dissected by marine erosion into a number of isolated segments, forms a headland known locally as the White Cow Rock. This name refers to a large mass of white vein-quartz, within the epidiorite, which simulates the form of a cow's head and encloses an "eye" of amphibole-biotite-skarn. A photograph of the White Cow Rock has already been published (Holmes and Reynolds, 1947, Fig. 2, p. 27), and a map of the White Cow epidiorite and its immediate environs appears in an earlier number of this *Magazine* (Reynolds and Holmes, 1954, Text-fig. 6, p. 424).

In the *Inishowen Memoir* of the old Geological Survey of Ireland, Nolan (1890, p. 29) records that the White Cow epidiorite encloses "fragments of gneiss . . . so highly altered as to pass in places into hornblende schist". Re-investigation has revealed that the "fragments of gneiss" have been evolved from thin sheets of kinematically and metasomatically transformed quartzite or micaschist that formed the soles of fold-thrusts with fold-axes plunging at 20° or more towards the north-west (Reynolds and Holmes, 1954, pp. 423–426). In illustration of this phenomenon a narrow sheet of progressively transformed quartzite, which can be seen to be continuous with the quartzite forming the floor of the epidiorite, has been selected for detailed investigation (Text-fig. 1). In the present paper, field and petro-chemical data are recorded concerning the kinematic and metasomatic transformation

of the quartzite to a " vein " of trondhjemite aplite, and of the adjoining epidiorite to biotite-epidiorite.

II. Field Evidence

An elevation of the south-eastern side of the largest erosion segment of the epidiorite (Text-fig. 1) is depicted in Text-fig. 2. The elevation

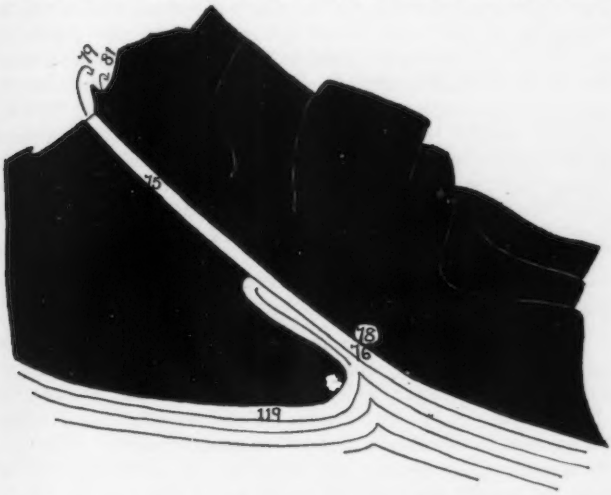


Text-fig. 1.—Plan of White Cow epidiorite (black) and a " vein " of transformed quartzite (TV) emplaced within it by fold-thrusting. VQ indicates vein quartz.

is essentially a strike section in which the epidiorite and the underlying quartzite dip north-westward, away from the observer, at an angle of about 55°. On the right-hand side of Text-fig. 2, a layer of quartzite forming the floor of the epidiorite has been upthrust towards the south-west, together with the overlying epidiorite, so that it now presents the appearance of a rheomorphic vein about 6 inches wide. The well-preserved form of the lower limb of the fold-thrust depicted in Text-fig. 2 and previously illustrated photographically (Reynolds and Holmes, 1954, Plate XI, Fig. 1.) provides conclusive evidence that

the movements responsible for the emplacement of the " vein " were glide movements of solid rocks rather than the flow of a melt or of a fluidized system.

The transformed quartzite of the " vein " occurs at the sole of the upper limb of the fold-thrust. In the elevation, Text-fig. 2, the fold-closure of this upper limb has been removed by denudation. Relatively leucocratic and more or less schistose hook-like antiforms, representing



TEXT-FIG. 2.—Elevation of the south-eastern side of the epidiorite of White Cow Rock (Text-fig. 1) showing the " vein " of trondhjemite aplite to be transformed quartzite at the sole of a fold-thrust. Numbers refer to samples. The width of the elevation is about 18 feet.

the upper limbs of fold-thrusts parallel to that under discussion are, however, well represented within the epidiorite. Narrow zones of epidiorite forming the foot- and hanging-walls of the " vein " are laminated parallel to its walls as a result of the glide-movements responsible for the emplacement of the " vein ". Conjointly, these two zones of laminated epidiorite are about three or four times as broad as the leucocratic " vein " that they margin. The zones of laminated epidiorite cut right through the exposed epidiorite, striking from 55° to 75° west of north, and dipping towards east of north parallel to the " vein " (Text-fig. 1). The leucocratic " vein ", however, peters out on the topographic surface, about 40 feet from the south-eastern margin of the epidiorite as depicted on the map (Text-fig. 1), probably as a

result of the north-westerly plunge of the axis of the fold-thrust under discussion. The vein-quartz of the " cow's head " occurs at a higher structural level than the leucocratic " vein ". The amphibole-biotite-skarn forming the " eye " of the " cow " is a transformed relic of the laminated epidiorite.

In the field it is apparent that the leucocratic " vein " is more melanocratic than the quartzite that constitutes the immediate floor of the epidiorite and of which it is the structural continuation (Text-fig. 2). Moreover, as the " vein " rises upwards through the epidiorite it assumes a more and more " igneous " appearance. In order to study the petro-chemical changes responsible for this transformation, specimens were collected at intervals both along the " vein " and the adjacent zones of laminated epidiorite. In particular two correlated pairs of specimens (76 and 78 ; 79 and 81) were collected from the " vein " and the adjacent laminated epidiorite for chemical investigation. The epidiorite (78, 81) of each of these pairs was taken from the upper limb of the fold-thrust as indicated in Text-figs. 1 and 2.

III Petro-chemical Evidence

Quartzite

The quartzite forming the immediate floor of the epidiorite sheet is composed essentially of quartz with a relatively small amount of feldspar. The feldspar, which is usually untwinned, is commonly albite (optically + ve with refractive indices lower than that of canada balsam, and 2V large, indicating low temperature albite). Texturally, the quartzite is fine-grained and schistose; the schistosity resulting from the parallel elongation of quartz grains (see Plate IX, Fig. 1). Epidote and chlorite are present in accessory amounts and form individuals about double the size of the quartz grains, the long axes of the epidotes and the cleavage of the chlorite usually being aligned parallel to the schistosity. Rarely, crystals of epidote and chlorite intersect the schistosity, and in such instances are evidently of relatively late crystallization.

A chemical analysis and the normative composition of the quartzite (119) from immediately beneath the epidiorite, as indicated in Text-fig. 2, are recorded under 1 in Tables 1 and 2 respectively.

The leucocratic " vein"

The leucocratic " vein " of Text-fig. 2, which is structurally continuous with the quartzite, gradually changes in mineral composition and grain-size as it rises upwards through the epidiorite. *Specimen 76*, collected near its base (Text-fig. 2) is composed of highly irregular crystals of variable grain-size. On average, the grain-size is about double that of the quartzite (119) described above. The most abundant

mineral is albite which occurs both as small untwinned grains, and as larger individuals of chess-board type (Plate IX, Fig. 2). The chlorite of the initial quartzite (119) is represented by biotite which forms relatively large individuals and commonly occurs as stellate aggregates ; the stellate form providing evidence that crystallization of biotite continued after movement had ceased. Basal sections of biotite have crenulate margins, as have the leucocratic constituents. Epidote, in accessory amount, builds larger crystals than those of the original quartzite ; other accessories include rare but relatively large grains of apatite, aggregates of sphene in which the individual crystals have wedgelike terminations, and rare stellate aggregates of acicular hastingsite.

A chemical analysis and norm of specimen 76, from the base of the leucocratic " vein ", are recorded under 2 in Tables 1 and 2 respectively. Chemically the rock is a basic trondhjemite (Goldschmidt, 1916). Comparison with Goldschmidt's many descriptions of trondhjemite reveals that specimen 76, and specimens 75 and 79 described below, are mineralogically most closely comparable with the layers of trondhjemite aplite in the *lit-par-lit* gneiss of the Stavanger area (Goldschmidt, 1921). The " vein " rock is adequately described as basic albite-trondhjemite aplite.

As compared with the quartzite no 1, Table 1, with which it is structurally continuous, the trondhjemite aplite (76) is considerably richer in Al_2O_3, Na_2O, CaO and to a lesser extent in iron oxides, MgO, H_2O, and K_2O ; correspondingly it is considerably poorer in SiO_2 than the initial quartzite (119).

Specimen 75, collected further along the leucocratic " vein ", i.e., from a higher topographic level, differs from 76 in that it is coarser grained and contains garnets and porphyroblasts of hastingsite (Z = bluish green, Y = yellowish green, X = pale straw colour). The rock, as is characteristic of many aplites, has a highly uneven grain-size, and the margins of the grains are crenulate. The hastingsite, which resembles that of the epidiorite in its optical properties (Holmes and Reynolds, 1947), forms relatively large porphyroblasts with a lace-like appearance. This appearance arises from the facts that the hastingsite encloses, in a sieve-like fashion, granules of albite and quartz like those of the rock matrix, whilst its highly irregular margins extend in cusp-like form along the junctions of the matrix crystals (Plate IX, Fig. 2). The unbroken lace-like form of the hastingsite indicates that it crystallized after movement had ceased. Some of the larger garnets, like the hastingsite, form sieved porphyroblasts, whilst the smaller garnets show some tendency to develop crystal form.

As in specimen 76, aggregates of biotite and sphene are present, and black iron ore forms an additional accessory mineral. The smaller

grains of albite are characteristically untwinned and, as in 76, the relatively large albites are of chess-board type. Streams of minute inclusions traverse the rock regardless of the grain-boundaries.

Specimen 79 represents the " tip " of the trondhjemite aplite " vein ", i.e., it was collected from the point where the " vein " peters out on the

TABLE 1

	Chemical Analyses			*Analyst :*	*Agnes Gibbs*	
	1	2	3	4	5	6
SiO_2	86·68	64·86	64·74	47·32	42·03	49·11
Al_2O_3	6·78	17·69	16·11	17·39	17·48	14·56
Fe_2O_3	0·54	1·57	1·60	2·36	4·12	3·70
FeO	1·21	3·09	4·84	10·91	11·18	10·91
MgO	0·20	0·66	1·19	4·78	5·18	4·93
CaO	1·22	4·01	3·70	10·08	9·24	8·94
Na_2O	1·56	5·34	5·25	2·73	2·20	1·73
K_2O	0·73	1·00	0·67	0·54	2·19	0·94
H_2O+	0·43	0·86	0·75	0·96	2·40	1·94
H_2O-	0·09	0·11	0·08	0·05	0·01	0·12
CO_2	trace	none	none	trace	trace	trace
TiO_2	0·37	0·62	0·79	2·68	2·27	2·35
P_2O_5	0·15	trace	0·75	trace	0·98	0·25
S	—	—	0·02	—	0·14	—
MnO	0·02	0·11	0·07	0·20	0·21	0·20
BaO	trace	none	trace	0·01	0·30	0·06
	99·98	99·92	100·26	100·01	100·02	99·82
Less O			·01		·06	
			100·25		99·96	

1. Quartzite, no. 119, forming the immediate floor of the epidiorite sheet of White Cow Rock, Pebble Strand, Malin Head.
2. Trondhjemite aplite " vein ", no. 76, structurally continuous with the quartzite represented by no. 119, recorded under 1 above. Fluorine was found to be absent.
3. Trondhjemite aplite " vein ", no. 79, in continuity with and from a higher structural level than no. 76, recorded under 2 above.
4. Epidiorite, no. 94, from the main epidiorite sheet of White Cow Rock.
5. Biotite-epidiorite, no. 78, adjoining the trondhjemite aplite vein no. 76, recorded under 2 above. The total includes Cr_2O_3 = trace, V_2O_3 = 0·07, NiO = 0·02.
6. Biotite-epidiorite, no. 81, adjoining the " tip " of the trondhjemite aplite " vein " no. 79, recorded under 3 above. The total includes NiO = 0·08.

topographic surface as a result of the north-westerly plunge of the axis of the fold-thrust. Mineralogically and texturally it resembles 75, described above. It differs from it, however, in that patches of the albite and quartz have a greater grain-size. Evidence that the relatively coarse-grained patches are of late crystallization is provided by relics of the initial fine-grained rock included within them. As in 75, lace-like porphyroblasts of hastingsite are present (Plate IX, Fig. 3). The

garnets are larger than in 75, whilst epidote occurs as aggregates in which the component grains have a smaller grain-size than the hastingsite, biotite, and garnet. Biotite forms aggregates as in 75 and 76, but some individuals of biotite differ from those in the specimens described above, in that they are interleaved with chlorite parallel to their cleavage. The boundaries between such biotite and chlorite are sharp and there is no appearance to suggest that the chlorite is secondary after biotite.

TABLE 2

Normative Compositions of Analyses in Table 1

	1	2	3	4	5	6
Q	71·72	18·82	21·22	—	—	6·36
C	1·50	0·54	1·64	—	—	—
Or	4·34	5·90	3·95	3·17	0·67	5·57
Ab	13·21	45·14	44·31	23·07	18·62	14·63
An	5·15	19·89	13·94	33·65	31·49	29·18
Lc	—	—	—	—	9·64	—
Di $CaSiO_3$	—	—	—	6·84	3·82	5·70
Di $MgSiO_3$	—	—	—	3·12	1·85	2·67
Di $FeSiO_3$	—	—	—	3·67	1·91	2·97
Hy $MgSiO_3$	0·50	1·65	2·96	4·98	—	9·61
Hy $FeSiO_3$	1·20	3·56	5·80	5·87	—	10·64
Ol Mg_2SiO_4	—	—	—	2·68	7·74	—
Ol Fe_2SiO_4	—	—	—	3·47	8·97	—
Mt	0·79	2·27	2·32	3·42	5·96	5·38
Il	0·70	1·18	1·50	5·08	4·31	4·46
Ap	0·37	—	1·78	—	2·32	0·61
Pr	—	—	0·01	—	0·26	—
water	0·52	0·97	0·83	1·01	2·41	2·06
	100·00	99·92	100·26	100·03	99·97	99·84

The chemical analysis and norm of specimen 79 are recorded under 3 in Tables 1 and 2 respectively. The rock is a little less rich in alkalis and alumina than 76, and correspondingly a little richer in iron oxides, MgO, and TiO_2. Within the " vein " itself there is, therefore, a culmination of alkalis and alumina near its base and a culmination of mafic constituents towards its " tip ".

Epidiorite

A petrochemical description of the main epidiorite was published fourteen years ago (Holmes and Reynolds, 1947). Here, it only requires to be stated in summary that the rock is markedly foliated being composed of an alternation of narrow melanocratic and leucocratic bands. The former are rich in hastingsite, and contain some iron ore, sphene, a little epidote and occasional biotite ; whilst the latter are composed of

feldspar, small crystals of epidote and occasional garnet. Broadly speaking, the epidiorite has the chemical composition of olivine-basalt; compared with analysed olivine-basalts, however, it is richer in Al_2O_3 and poorer in MgO (Table 1, no. 4). In these differences from normal olivine-basalt it is comparable with the epidiorite described by Wiseman (1934, p. 378) from the garnet zone of South Knapdale, in the northern part of the Kintyre peninsula, Argyllshire.

Biotite-epidiorite

Adjacent to the " vein " of trondhjemite aplite, the epidiorite becomes more melanocratic and contains biotite in essential proportions. These basic selvages margining the " vein " are conjointly three or four times as wide as the " vein " itself. They are, moreover, highly laminated parallel to the " vein " as a result of the glide-movements responsible for the fold-thrust. On the map of the White Cow rocks (Holmes and Reynolds, 1947, Fig. 3, p. 29) these laminated melanocratic zones were originally described as " crush belts ". The expression " shear zone " might be more appropriate, but with the proviso that recrystallization outlasted movement; in their present form the selvages of melanocratic epidiorite are neither " crushed " nor " sheared ".

The selvages of melanocratic epidiorite adjoining the " vein " are coarser grained than the normal epidiorite. Within them hastingsite, with its prisms aligned parallel to the lamination, is the most abundant mineral. Biotite, which is not only more abundant but builds larger crystals than in the normal epidiorite, increases in amount downwards, being most abundant adjacent to the base of the " vein ". Commonly it is interleaved with chlorite, as in 79. Other constituents of the biotite-epidiorite are aggregates of sphene aligned parallel to the lamination, black iron ore, rare small garnets and sparse granules of epidote. The leucocratic constituents are finer grained and less abundant than in the normal epidiorite. They include quartz and untwinned oligoclase.

The chemical analyses of specimens 78 and 81, respectively adjacent to the base and " tip " of the " vein ", are recorded under 5 and 6 in Table 1. The normative compositions of specimens 78 and 81 are recorded under 5 and 6 in Table 2. In comparison with the normal epidiorite recorded under 4, Table 1, both specimens of biotite-epidiorite are richer in K_2O, iron oxides and H_2O, and correspondingly poorer in in Na_2O, CaO, and TiO_2. The various constituents are, moreover, distributed differentially within the selvage of biotite-epidiorite. Adjacent to the " tip " of the " vein " the biotite-epidiorite is richer in SiO_2 and correspondingly poorer in Al_2O_3, Na_2O, K_2O, CaO, iron oxides and H_2O than is the biotite-epidiorite adjacent to the base of the " vein ".

IV Petrogenesis

The petrogenetic problems arising out of this field and petrochemical study relate to the evolution of :—(*a*) trondhjemite aplite from a narrow layer of quartzite that formed the sole of a fold-thrust, and (*b*) melanocratic selvages of biotite-epidiorite from the main epidiorite adjacent to the trondhjemite aplite. In other words, by what means has the quartzite been granitized and the adjacent basic fronts evolved from epidiorite ?

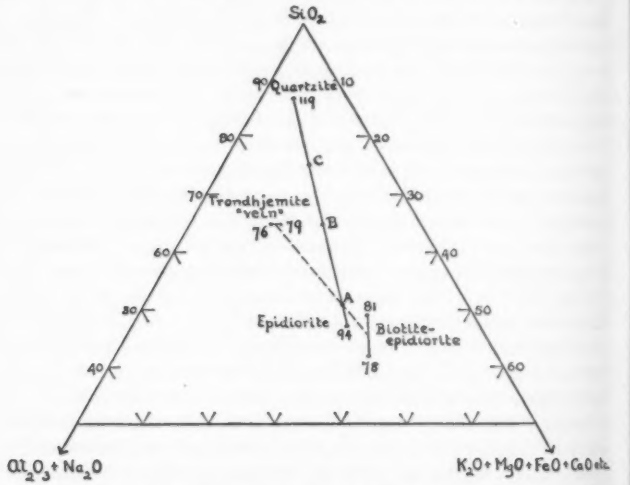
There are still many petrologists who regard it as axiomatic that the agent responsible for granitization is granite magma. This causal hypothesis is not applicable to the granitization of quartzite at the White Cow Rock for the very strong reason that there is no granitic rock in the vicinity, apart from the " vein " of trondhjemite aplite itself, which is seen to be structurally rooted in quartzite (Text-fig. 2).

A possible explanation of the trondhjemite " vein " and its basic selvages is that they have been formed respectively from the pre-existing quartzite and epidiorite as a result of *metamorphic differentiation.* The expression *metamorphic diffusion* has been used to describe the process responsible for observed changes of composition (metamorphic differentiation) in rocks within metamorphic terrains ever since Stillwell (1911–18) first introduced these terms in his study of the regional metamorphic rocks of Adelie Land, Antarctica. Eskola (1932, 1933) has done much to make the concept of metamorphic diffusion familiar to geologists. In particular, he (1934) summarized some of the early experiments which led to the recognition that shearing, i.e. differential movement in solids, promotes diffusion and chemical reactions. Highly relevant to the present investigation is Sven Gavelin's (1960) recent suggestion that the feldspathization of kinematically deformed quartzites, adjacent to metabasites in the Västervik area, Sweden, has been caused by some kind of metamorphic exchange. As at the White Cow Rock, this feldspathization is not associated with intrusions of granite, granodiorite or tonalite, and so cannot be explained as a result of transfer of matter from granitic magma.

The concept of metamorphic differentiation is a highly complex one, for changes of chemical and mineralogical composition may not depend on metamorphic diffusion alone, but also on mechanical mixing. Hitherto the details relating to these two aspects of the subject have tended to be studied separately, the chemical aspect by petrologists and the kinematics by tectonicians. In the present study of the White Cow Rock, the chemical data relate to a structural form (fold-thrust, Text-fig. 2) that in itself provides evidence of the relevant kinematics. It is, therefore, possible to discover whether a redistribution of the materials of the initial quartzite and epidiorite is adequate to account for the trondhjemite and biotite-epidiorite, and if so to decide

whether such a redistribution of materials is best explained by :— (*a*) mechanical mixing, or (*b*) metamorphic diffusion, or (*c*) by some combination of these two processes.

First, then, it has to be decided whether or not the chemical compositions of the initial quartzite and epidiorite are adequate to account for the transformation of quartzite to trondhjemite and the correlated enrichment of the adjacent epidiorite in biotite, or whether introduction of material from an external source was necessary. On Text-fig. 3, the



TEXT-FIG. 3.—Diagrammatic illustration of the chemical relationships between the analysed rocks. The letters, A, B, and C, refer to calculated chemical compositions recorded in Table 3, and discussed under petrogenesis.

chemical analyses (Table 1) of all the rocks concerned are plotted. This simple type of triangular diagram has been chosen because :—

(1) it expresses the total chemical compositions of the rocks, and (2) it adequately separates the initial quartzite and epidiorite. In constructing the diagram the potash has been added to the cafemics rather than to soda and alumina, because the biotite-epidiorite is richer in potash than the trondhjemite. A diagram with potash added to soda and alumina has also been drawn, and found to give similar results.

In Text-fig. 3 the line joining quartzite (119) and epidiorite (94) is intersected at A by a line joining the average of the two analyses of

trondhjemite with the average of the two analyses of biotite-epidiorite. This makes it clear that a mixture of quartzite and epidiorite of composition A is equivalent to an average mixture of the trondhjemite and biotite-epidiorite, with the proviso that where a corner of the triangle represents more than one oxide, there may be a change in the relative proportions of these oxides. The composition of A (Text-fig. 3) can, however, be calculated from the initial quartzite (119) and epidiorite (94) by combining their chemical analyses (nos. 1 and 4, Table 1) in the

TABLE 3

	7	8	9	10
SiO_2	50·90	50·66	65·03	73·56
Al_2O_3	16·43	16·25	12·62	10·32
Fe_2O_3	2·19	3·29	1·54	1·15
FeO	10·03	9·14	6·55	4·44
MgO	4·36	3·97	2·72	1·73
CaO	9·27	7·76	6·09	4·17
Na_2O	2·62	2·85	2·20	1·95
K_2O	0·56	1·38	0·63	0·67
H_2O+	0·91	1·81	0·72	0·61
H_2O-	0·05	0·08	0·07	0·08
TiO_2	2·47	1·89	1·64	1·14
P_2O_5	0·01	0·56	0·07	0·10
MnO	0·18	0·22	0·12	0·07
BaO	0·01	0·13	0·01	trace
	99·99	99·99	100·01	99·99

7. Composition of A, Text-fig. 3, calculated from the initial quartzite (119) and epidiorite (94).
8. Composition of A, Text-fig. 3, calculated from the average compositions of the trondhjemite (76, 79) and biotite-epidiorite (78, 81).
9. Composition of B, Text-fig. 3, a mixture of 45 per cent quartzite (119) and 55 per cent epidiorite (94).
10. Composition of C, Text-fig. 3, a mixture composed approximately of two-thirds quartzite (119) and one-third epidiorite (94).

respective proportions indicated by the position of A. The composition of A, calculated in this way, is recorded under 7 in Table 3. The composition of A, calculated in a similar way from the average of the two analyses of the trondhjemite and the average of the two analyses of the biotite-epidiorite is recorded under 8 in Table 3. The two calculated compositions of A (7 and 8, Table 2) are remarkably similar and from them it can be concluded that a redistribution of the materials of the initial quartzite and epidiorite, represented by A on Text-fig. 3, is chemically adequate to account for the average composition of the trondhjemite and biotite-epidiorite, with but minor additions of K_2O, H_2O, and P_2O_5, and a small loss of CaO.

An attempt is now made to assess the parts played by mechanical

mixing and metamorphic inter-diffusion respectively in the transformation of quartzite to trondhjemite at the sole of the thrust. The composition of any mechanical mixture of quartzite and epidiorite that may have resulted from shear movements associated with the evolution of the fold-thrust must lie somewhere along the line connecting the compositions of the quartzite (119) and epidiorite (94) in Text-fig. 3. Since the analyses of trondhjemite do not lie on this line, the trondhjemite cannot be accounted for by recrystallization of a mechanical mixture of quartzite and epidiorite without metamorphic inter-diffusion.

On the other hand, reference to the relevant chemical analyses in Table 1 reveals that if it be supposed that the trondhjemite was formed from the quartzite by metamorphic inter-diffusion of the materials of the initial quartzite and epidiorite, without mechanical mixing, then it must be supposed that Al_2O_3, iron oxides, MgO, CaO, Na_2O, H_2O, TiO_2 and MnO have been added to the quartzite and considerable amounts of SiO_2 lost. If metamorphic inter-diffusion was the process concerned then these gains and losses should be balanced respectively by the average decreases and increases in the chemical analyses of the biotite-epidiorite in comparison with the initial epidiorite. They are not.

It may be concluded that the transformation to trondhjemite of quartzite, at the sole of the fold-thrust, involved both mechanical mixing and metamorphic inter-diffusion. Indeed the shearing movements, by virtue of which the fold-thrust was formed, would inevitably cause a certain amount of mechanical mixing of the quartzite with the epidiorite over which it rode (Text-fig. 2). Differential movement between the quartzite sole and the overlying epidiorite might cause additional mechanical mixing. Furthermore, metamorphic inter-diffusion may have been promoted by the shearing stress which would inevitably distort and probably to some extent destroy crystal lattices.

The composition of any mechanical mixture of quartzite and epidiorite that may have resulted from shear-movements associated with the evolution of the fold-thrust must lie somewhere along the line connecting the compositions of the quartzite (119) and epidiorite (94) in Text-fig. 3. This composition can only be determined within limits. One of these limits, B on Text-fig. 3, is fixed by the silica percentage of the trondhjemite ; if B lay nearer to the epidiorite, then SiO_2 would have to be added from an external source. The other limit C is fixed by the iron oxides (i.e., total Fe) of the mixture. The transformation of the epidiorite to biotite-epidiorite, adjacent to the trondhjemite " vein ", involved addition of iron oxide (compare nos. 5 and 6 with 4, Table 1). This implies that the mechanical mixture of quartzite and epidiorite contained more iron oxide than the trondhjemite to which it was eventually transformed, i.e., that it contained sufficient epidiorite to

account for this excess iron oxide, and that the excess was lost to the wall-rocks by metamorphic diffusion. If C lay nearer to the quartzite there would be no excess of iron oxide over that in the trondhjemite to account for that added to the epidiorite wall-rocks.

The two limiting compositions B and C (Text-fig. 3) of the mechanical mixture of quartzite and epidiorite are recorded under 9 and 10 in Table 3. Comparsion of both these compositions with those of the trondhjemites (2 and 3, Table 1) reveal that :—(*a*) a considerable amount of Al_2O_3 and Na_2O and a small amount of K_2O would have to be added to the mechanical mixture within the limits B and C to convert it to trondhjemite, and (*b*) smaller amounts of iron oxide, MgO and CaO would have to be lost. The Al_2O_3 and Na_2O added to the mechanical mixture to make trondhjemite are balanced by corresponding losses from the adjacent epidiorite during its transformation to biotite-epidiorite. On the other hand iron oxide and MgO lost from the mechanical mixture are balanced by increase of these materials within the selvages of biotite-epidiorite in comparison with the initial epidiorite.

Silica lost from the mechanical mixture in its transformation to trondhjemite is not balanced by increased silica in the biotite-epidiorite selvages. That it has travelled upwards, along the " vein " is evidenced by the " cow's head " of vein quartz. Similarly some lime, as found in the calculation of composition A (Table 2, no. 1), is lost from the system.

V Conclusion

It can be concluded that the trondhjemite " vein " was formed by the combined processes of (*a*) mechanical mixing of quartzite, at the sole of the thrust, with epidiorite in respective percentages by weight of about sixty and forty, and (*b*) metamorphic diffusion. With but a small addition of K_2O the metamorphic diffusion can be adequately accounted for by local interchanges between the mechanical mixture at the sole of the thrust and the epidiorite wall rock. Alumina and soda, lost from the epidiorite wall-rock, correspond with the necessary additions to the mechanical mixture to form the trondhjemite of the " vein ". On the other hand, iron oxide (total iron) and MgO lost from the mechanical mixture correspond with the introductions into the wall-rock of the trondhjemite " vein " where, with the addition of a little K_2O from an external source, selvages of biotite-epidiorite were formed from the epidiorite.

Metamorphic differentiation also proceeded in a vertical sense, for both the trondhjemite and the biotite-epidiorite selvages are richer in alkalis near the base of the " vein" than at its " tip ". The trondhjemite " vein " is poorer in alkalis, relative to silica, near its " tip " and richer

in iron oxides and magnesia. On the other hand, the biotite-epidiorite selvage of the upper limb of the fold is enriched in silica near the " tip " of the " vein ".

Increases and decreases of the various constituents have been discussed in terms of oxides, as they are expressed in the chemical analyses of Table 1, without prejudice as to the actual mode of migration.

Some petrologists may question the desirability of using Goldschmidt's term trondhjemite to describe a rock that is known to have had a metamorphic origin. In conclusion it is, therefore, important to refer to Goldschmidt's own views on this subject. In a letter to Dr. Donald B. McIntyre (1947, pp. 62–3), concerning the use of a rock name introduced by Goldschmidt in describing the rocks of the Opdal-Trondhjem area, Professor Goldschmidt wrote :— " At the present time when the truly magmatic or metasomatic status of so many crystalline rocks is under discussion, we cannot afford to create two different sets of names ".

VI Acknowledgments

The chemical analyses that form an essential part of the paper were made by the late Lady Gibbs, in Dr. H. F. Harwood's laboratory at Imperial College, a few years before her untimely death in 1939. For these and many other chemical analyses I owe a great debt of gratitude to my old friend and valued collaborator.

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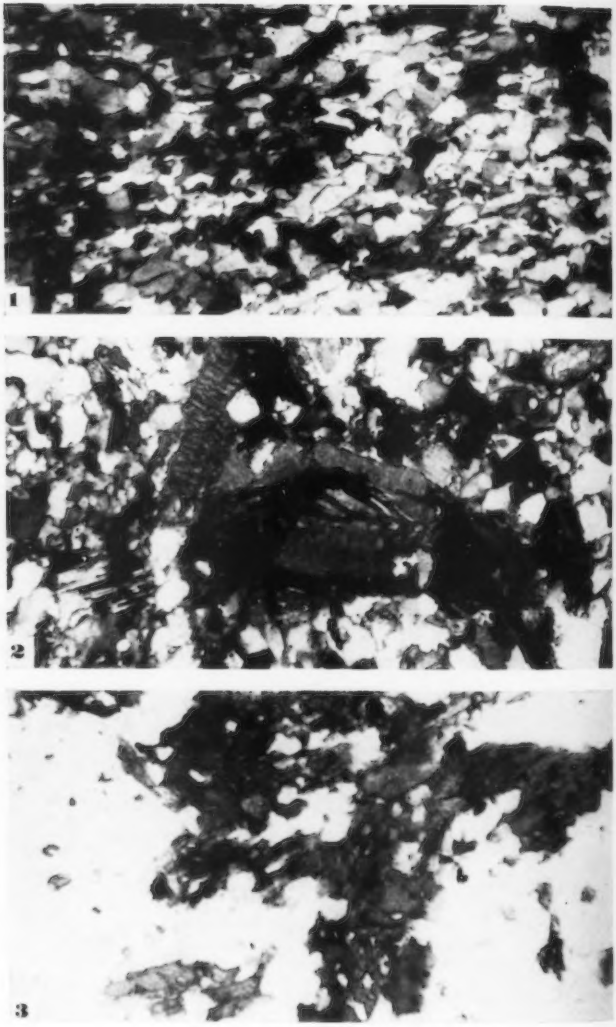

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EXPLANATION OF PLATE IX

FIG. 1.—Quartzite, no. 119, forming the immediate floor of the epidiorite sheet of White Cow Rock, Pebble Strand, Malin Head. Nicols crossed, × 44. Photomicrograph by Miss Jean Tarrant.

FIG. 2.—Albite-trondhjemite aplite, no. 76, structurally continuous with the quartzite of Fig. 1 above. The photomicrograph shows an aggregate of biotite crystals the stellate arrangement of which indicates growth after movement had ceased. Towards the lower left-hand corner is a crystal of chess-board albite. Nicols crossed, × 44. Photomicrograph by Miss Jean Tarrant.

FIG. 3.—Albite-trondhjemite aplite, no. 79, in structural continuity with no. 76, Fig. 2 above. The photomicrograph shows porphyroblasts of hastingsite which enclose granules of quartz and albite, like those of the rock-matrix. The irregular margins of the hastingsite extend in cusp-like form along the junctions of matrix crystals. Ordinary light, × 44. Photomicrograph by Miss Jean Tarrant.

Some Namurian Conodonts from North Staffordshire

By A. C. Higgins

(PLATES X–XII)

Abstract

The paper describes a conodont fauna from the Namurian of North Staffordshire. Seventeen genera and thirty-four species are identified in the fauna, of which *Hibbardella fragilis*, *Hibbardella pennata*, *Lambdagnathus macrodentata*, *Mestognathus bipluti*, *Neoprioniodus spathatus*, and *Subbryantodus subaequalis* are new species. The composition of the fauna and its correlation with similar faunas in the United States of America and Germany are discussed.

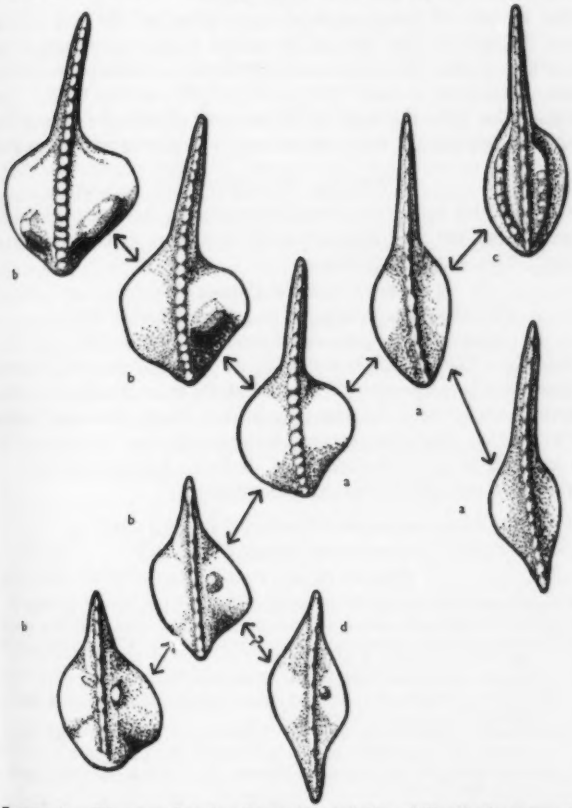
Introduction

The last forty years have seen the publication of many papers, particularly by American and German workers, on conodont faunas from the Lower Carboniferous and that part of the Upper Carboniferous equivalent to the American Pennsylvanian. Such work as that of Branson and Mehl (1933–34, 1941), Hass (1953, 1959), Bischoff (1957) and Voges (1959), in particular has provided considerably detailed knowledge of the faunas from the American and German Lower Carboniferous sequences. At the same time knowledge of Upper Carboniferous forms from the American successions has resulted from the works of Ellison (1941), and Gunnell (1931, 1933). Such works amongst others have amply demonstrated the potential value of conodonts as useful index fossils in problems of local stratigraphy, and at the same time have indicated their value in intercontinental correlation problems. The universal distribution of these remains and their preservation in a variety of rock types suggests they may be particularly appropriate in long distance correlation when further work and a greater number of records concerning their distribution is available.

The present paper constitutes one of the first detailed records of a conodont fauna from the British Carboniferous and more precisely from a well-defined stratigraphical horizon, namely the *Eumorphoceras* aff. *pseudobilingue* marine band. This horizon is of particular interest in that it represents an horizon which is near the Upper and Lower Carboniferous junction as based on goniatite zones. Few detailed records of conodonts from beds equivalent in age to Namurian are known from anywhere in the world. Only Ellison (1941), Branson and Mehl (1941), Youngquist and Miller (1949), Elias (1956) and Rexroad (1957, 1958) have given detailed descriptions of conodonts from these beds.

Stratigraphic Sequence, Locality, and Techniques

The conodont fauna described in this paper was obtained from the *Eumorphoceras* aff. *pseudobilingue* marine band exposed in a railway



Text-fig. 1.—Diagram showing the major varieties of *Gnathodus commutatus* and the possible relationship of this species to *Spathognathodus* sp., a. *Gnathodus commutatus commutatus*, b. *Gnathodus commutatus nodosus*, c. *Gnathodus commutatus homopunctatus*, d. *Spathognathodus nodosus*, × 15.

cutting 250 yards north of the village of Cauldon, North Staffordshire (076496).

The conodonts were present in abundance (500–700 specimens per kilogram of rock) and occurred in a thin band of limestone 1½ inches

thick which passed through a line of bullions. In this rock the conodonts were associated with recognizable goniatites, identified by Bisat (Ludford, 1950), and with lamellibranchs, gastropod spat, ostracods, fish teeth and spines, and other microfossils.

The samples of limestone band were taken at 100-yard intervals along the section, the amount of sample being determined by the availability of material. The maximum amount available was approximately 1 kilogram of rock from any one place in the band. These samples were then digested in 15 per cent acetic acid, the residue washed and sieved, and the organic remains recovered by hand picking.

Systematic Palaeontology

All types and figured specimens illustrated in this paper are in the collection of the micropalaeontology laboratory, Department of Geology, University of Sheffield.

Genus *Gnathodus* Pander 1856

(Type species: *Gnathodus mosquensis* Pander 1956)

Gnathodus commutatus Branson and Mehl 1941

Remarks.—This species is very variable but the following constant features have been observed:—approximately symmetrical unit with a posterior convex cup and an anterior median blade continued posteriorly as a carina. Such features as ornamentation, size, and shape of the cup and the fusion and number of denticles are considered to be characters of ontogeny or infraspecific in value.

Gnathodus commutatus Branson and Mehl 1941, var. *commutatus* Bischoff 1957

Plate X, fig. 6; Text-fig. 1a

1941. *Spathognathodus commutatus* Branson and Mehl, p. 98, pl. 19, figs. 1-4.
1941. *Spathognathodus commutatus* Branson and Mehl, p. 172, pl. 5, figs. 19-22.
1941. *Spathognathodus commutatus* (Branson and Mehl). Ellison and Graves, pp. 3–4, pl. 2, figs. 4–6.
1953. *Gnathodus inornatus* Hass, p. 80, pl. 14, figs. 9–11.
1957. *Gnathodus commutatus* sub. sp. *commutatus* (Branson and Mehl). Bischoff, p. 22, pl. 14, figs. 2–15.
1957. *Gnathodus commutatus* sub. sp. *commutatus* (Branson and Mehl). Ziegler, in Flugel and Ziegler, p. 39, pl. 3, fig. 21.
1958. *Spathognathodus commutatus* (Branson and Mehl). Rexroad, p. 38, pl. 3, figs. 23–24.
1958. *Spathognathodus* cf. *commutatus* (Branson and Mehl). Rexroad, p. 26, pl. 6, fig. 8.
1959. *Gnathodus commutatus* sub. sp. *commutatus* (Branson and Mehl). Voges, p. 281.

Remarks.—This variant at the present horizon possesses typically an almost circular cup folded on its inner side, and a blade as long as the carina. Other specimens show a lengthened oval cup with a blade one and one-half times the length of the carina. These latter forms grade into *Gnathodus commutatus* var. *homopunctatus* and differ from

this variant in the lack of ornament on the cup. Figured specimen: E1.E7. Number of specimens: 181

Gnathodus commutatus Branson and Mehl 1941, var. *nodosus* Bischoff 1957

Plate X, figs. 7, 8; Text-fig. 1b

1957. *Gnathodus commutatus* sub. sp. *nodosus*, Bischoff, p. 3, pl. 4, figs. 12–3.
1957. *Gnathodus commutatus* sub. sp. *nodosus* (Bischoff). Ziegler, in Flugel and Ziegler, p. 40, pl. 3, fig. 4.

Diagnosis.—Bischoff's diagnosis (Bischoff, 1957): " A sub-species of *Gnathodus commutatus* with one strong, lengthened node on the oral surface of the inner and outer halves of the cup ", is here amended to: A variety of *Gnathodus commutatus* with a node or nodes, on the inner, or inner and outer sides of the cup.

Description.—The general outline is similar to the outline of *Gnathodus commutatus commutatus*. The cup is ornamented by a node or nodes on the inner or inner and outer sides of the cup: The nodes may be small and rounded or fused into a lengthened node extending anteriorly at approximately 45° to the unit axis. The longest diameter of the cup forms an acute angle with the inner anterior side of the unit axis, except in a few specimens (see Text-fig. 1). Figured specimens: E1.E8,E1.E9. Number of specimens: 278.

Genus *Hibbardella* Ulrich and Bassler 1926

(Type species: *Hibbardella angulata* Hinde 1879)

Hibbardella pennata sp. nov.

Plate XII, figs. 5, 6

Diagnosis.—A *Hibbardella* with a shallow, delicate anterior arch and long posterior bar.

Description.—Anterior arch shallow. The anterior faces and aboral edges of the two lateral limbs are convex and separated by a groove. Each lateral limb has four or more posteriorly curved denticles on each side of the large, laterally compressed main cusp.

The posterior bar is long, arched, bowed, and varies in height, being low at the anterior end but heightening to a point one-third from the posterior end and then decreasing to the extremity.

The narrow aboral edge has a median groove which expands into a small pit beneath the main cusp.

Comparisons.—This species is distinguished from *Hibbardella milleri* Rexroad, 1958 and *Hibbardella ortha* Rexroad, 1958 by its fragility and its shallow anterior bar.

Holotype.—E1.E2.12. Number of specimens: 10.

Hibbardella fragilis sp. nov.

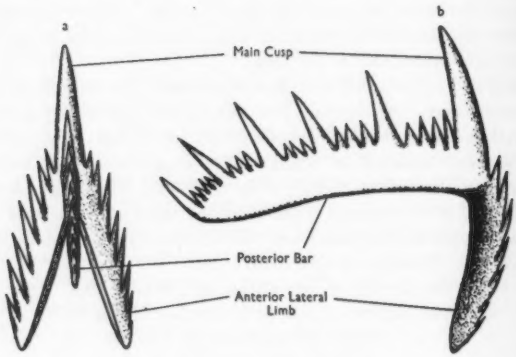
Plate XII, fig. 4; Text-fig. 2

Diagnosis.—A *Hibbardella* with fragile, long, lateral limbs, the anterior sides of which face inwards towards the long axis of the unit.

Description.—The unit is fragile. The anterior arch is deep and has two long, fragile, pointed, lateral limbs which are curved posteriorly and separated by an angle of 30°–40°. The anterior sides of the lateral limbs face inwards towards the long axis of the unit, and their outer edges bear six or more denticles. The laterally compressed main cusp is large and curved posteriorly.

The posterior bar is slightly longer than the lateral limbs and its oral edge bears numerous denticles usually of two sizes.

The aboral edge of the posterior bar and lateral limbs is thin and



TEXT-FIG. 2.—*Hibbardella fragilis* n. sp. a. Posterior View. b. Lateral View. × 15.

medianly grooved, the groove expanding into a small triangular pit beneath the main cusp.

Comparisons.—This species differs from *Hibbardella pennata* sp. nov. in possessing long lateral limbs and from *Hibbardella milleri* Rexroad, 1958 and *Hibbardella ortha* Rexroad, 1958 in that the anterior sides of the lateral limbs face inwards towards the long axis of the unit.

Holotype.—E1.E2.11. Number of specimens: 8.

Genus *Lambdagnathus* Rexroad 1958

(Type species: *Lambdagnathus fragilidens* Rexroad 1958)

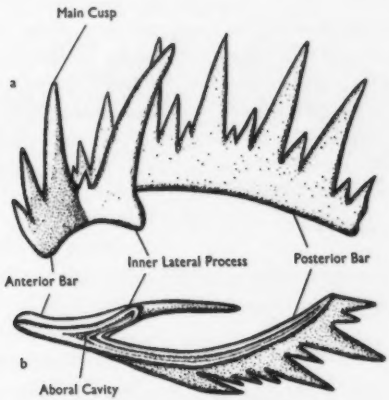
Lambdagnathus macrodentata sp. nov.

Plate XII, figs. 1–3; Text-fig. 3

Diagnosis.—A *Lambdagnathus* with the posterior bar the longest process on the unit; the inner lateral process of variable length and anterior bar the shortest process on the unit.

Description.—Oral View: Unit fragile. The inner lateral process is one-quarter to two-thirds the length of the posterior bar and the two processes are approximately 30° apart. The anterior bar is short, up to one-third the length of the inner lateral process, and it often bisects the angle of the other two processes.

Lateral view.—Posterior bar arched, is deepest near the midlength, and has posteriorly inclined and laterally compressed denticles which are largest near the midlength. The inner lateral process is slightly



TEXT-FIG. 3.—*Lambdagnathus macrodentata* n. sp. a. Inner Lateral View. b. Aboral View. × 35.

arched, and has denticles, two to five in number, which are usually small but with one large denticle expanded at its base on the inner side. The main cusp is triangular in cross-section at the base, and occurs at the junction of the three processes; it is as large as the next adjacent large denticle on the posterior bar. The anterior bar is as deep as the posterior bar, is downcurved slightly and bears two to four minor denticles.

Aboral view.—The aboral edge is thin, bevelled and striated, with a small triangular pit beneath the main cusp which is continued along the processes as a groove and expanded beneath the large denticle on the inner lateral process.

Comparison.—This species differs from *Lambdagnathus fragilidens* (Rexroad, 1958) in having a short anterior bar and inner lateral process and a long posterior bar.

Holotype.—E1.E2.3. Number of specimens: 45.

Genus *Mestognathus* Bischoff 1957

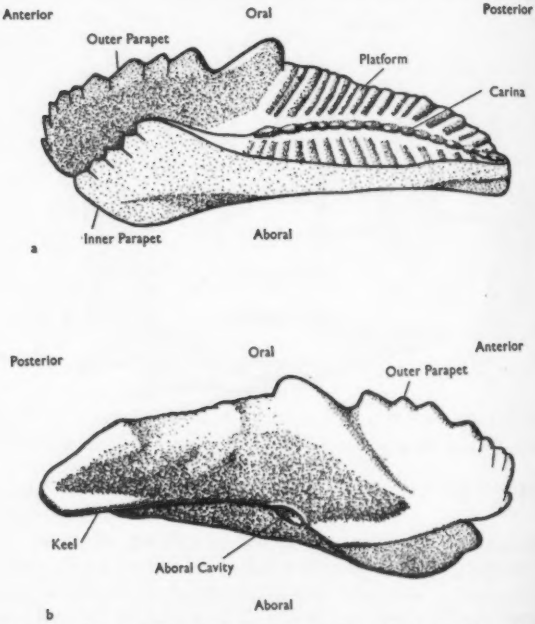
(Type species: *Mestognathus beckmani* Bischoff 1957)

Mestognathus bipluti sp. nov.

Plate X, figs. 1, 2; Text-fig. 4

Diagnosis.—A *Mestognathus* with two denticulate parapets, inner and outer, at the anterior end of the unit.

Description.—Oral view: Unit approximately triangular and pointed



TEXT-FIG. 4.—*Mestognathus bipluti* n. sp. a. Inner Lateral View. b. Outer Aboral View. × 15.

at the posterior end. A triangular platform, flat or slightly concave occupies the posterior half of the unit and is ornamented with strong ridges which degenerate towards the carina. The carina is often weak; it occupies a median position at the posterior end, but curves inwards at the anterior end and joins the inner parapet. The anterior end consists of two parapets separated by a deep trough. The larger outer parapet is half the length of the unit, whilst the smaller inner parapet is one quarter the length of the unit but it may extend further anteriorly.

Outer lateral view.—Unit arched. The oral surface of the platform is longitudinally convex and from it the outer parapet rises steeply. The outer parapet has numerous denticles decreasing in size anteriorly from the very large posterior one.

Inner lateral view.—The inner parapet is smaller than the outer parapet but it may extend farther anteriorly; its oral surface rises gently from the platform, and its denticles, usually six in number, are inclined posteriorly at 45°. The aboral edge of the inner parapet is usually parallel to the oral margin.

Aboral view.—The aboral side is convex, keeled for the posterior two-thirds of its length, and bifurcates one-third from the anterior end. Growth lines are present on the aboral surface. The pulp cavity is small, oval and continued as a groove along the keel.

Comparison.—This species differs from *Mestognathus beckmani* Bischoff in possessing a well-developed inner parapet.

Remarks.—*Mestognathus bipluti* appears to be unpaired since the curvature of individual specimens is independent of the inner and outer sides as defined on the size of the parapets. This suggests either:—

(i) They were not paired.

(ii) There were considerable differences between right and left forms.

(iii) There were differences in the direction of curvature between different pairs.

Holotype.—E1.E.1. Number of specimens: 25.

Genus *Neoprioniodus* Rhodes and Müller 1956

(Type species: *Prioniodus conjunctus* Gunnell 1931)

Neoprioniodus spathatus sp. nov.

Plate XI, figs. 2, 4; Text-fig. 5

Diagnosis.—A *Neoprioniodus* with a spatulate anticusp, and a bevelled inner aboral edge.

Description.—Unit slightly arched, bowed, and longitudinally concave on its inner side. The posterior bar is long, thin, gently arched, and bowed. The denticles, ten to fifteen in number, are pointed and decrease in size posteriorly. The main cusp is large, laterally compressed, and is much higher and wider than the remaining denticles. It is inclined anteriorly in relation to the posterior bar, and slightly twisted at the distal end.

The anticusp is large, up to one-quarter the length of the main cusp, is spatulate and has convex posterior and concave anterior edges. Its anterior edge bears one to five small denticles, but typically it bears one inwardly inclined denticle.

The aboral inner edge is bevelled. The aboral cavity is small and is continued as a median groove to the extremities of the unit.

Comparisons.—This species differs from *Neoprioniodus scitulus* Branson and Mehl, 1941 in possessing a spatulate anticusp and anterior denticles, but it is possible that a large collection would reveal gradational forms between these two species.

Holotype.—EI.E.22. Number of specimens: 71.

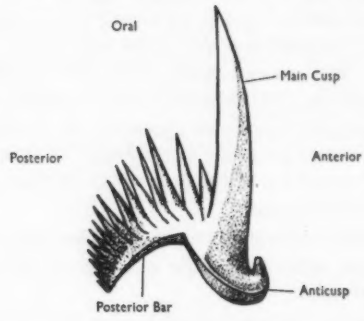
Genus *Spathognathodus* Branson and Mehl 1941

(Type species: *Spathodus primus* Branson and Mehl 1933)

Spathognathodus sp.

Plate X, fig. 10; Text-fig. 1d

Description.—*Oral view*: Axis almost straight, slightly curved at the



TEXT-FIG. 5.—*Neoprioniodus spathatus* n. sp. Lateral View. × 15.

anterior end. The denticles are laterally compressed but slightly thickened in the middle third of unit. The cup is almost symmetrically oval with the inner side slightly smaller than the outer; it extends almost the whole length of the unit and carries a small node on its inner side.

Lateral view.—The oral edge rises in a curving line from the posterior end. The anteriormost denticles are twice the length and width of the remaining denticles.

Aboral view.—The cup is deepest one-third of the distance from the anterior end.

Remarks.—This single specimen is referred to *Spathognathodus* because of its medially situated cup, but it is probably related to *Gnathodus commutatus nodosus* as suggested in Text-fig. 1. Figured specimen: E1.B10. Number of specimens: .

Genus *Subbryantodus* Branson and Mehl 1933

(Type species: *Subbryantodus arcuatus* Branson and Mehl 1933)

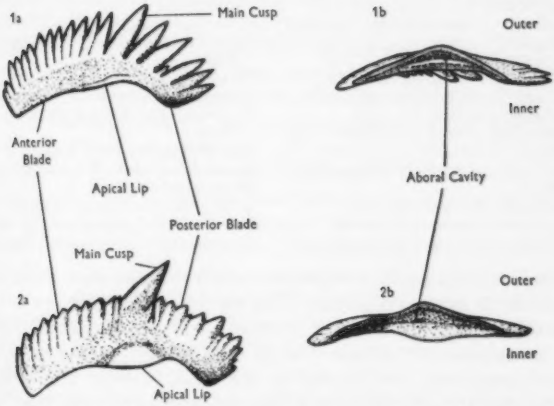
Subbryantodus subaequalis sp. nov.

Plate XII, fig. 15; Text-fig. 6

Diagnosis.—A *Subbryantodus* with the posterior blade denticles discrete and as large as, or slightly smaller than, the main cusp.

Description.—Unit broadest beneath main cusp, strongly arched, bowed, and compressed.

The anterior blade is slightly longer and higher than the posterior blade and downcurved about 45° to it. The denticles of the anterior blade are small, crowded, and curved and inclined in a posterior direction whereas the denticles of the posterior blade are higher and wider and decrease in size posteriorly. They are discrete, and as high or slightly shorter than the main cusp. The suppression of denticles is



TEXT-FIG. 6.—1. *Subbryantodus subaequalis* n. sp. a. Lateral View, b. Aboral View. 2. *Subbryantodus stipans*. a. Lateral View, b. Aboral View. × 15.

common. The distal end of the posterior blade is strongly bowed and incurved. The main cusp is pointed and inclined posteriorly.

The apical lip is small and extends one-third the length of the unit.

The aboral side is excavated and has thick lips and a small pit which is continued as a groove to the blade extremities.

Comparisons.—This species differs from *Subbryantodus stipans* Rexroad, 1957 in the presence of discrete, large denticles on the posterior blade.

Holotype.—E1.E.19. Number of specimens: 45.

DISCUSSION OF SOME ASPECTS OF THE CONODONT FAUNA

The abundance and variety of the conodont fauna represented in the *E.* aff. *pseudobilingue* Marine Band is shown in the species list which contains seventeen genera and thirty-four species. Of the species six

are new, *Hibbardella fragilis*, *Hibbardella pennata*, *Lambdagnathus macrodentata*, *Mestognathus bipluti*, *Neoprioniodus spathatus*, and *Subbryantodus subaequalis*. The following is a complete list of conodont species isolated from this band:—

Angulodus walrathi Hibbard
Cavusgnathus unicornis Youngquist and Miller
Geniculatus claviger Hass
Gnathodus bilineatus (Roundy)
Gnathodus commutatus commutatus Branson and Mehl
Gnathodus commutatus homopunctatus Bischoff
Gnathodus commutatus nodosus Bischoff
Gnathodus girtyi Hass
Hibbardella fragilis sp. nov.
Hibbardella milleri Rexroad
Hibbardella pennata sp. nov.
Hindeodella brevis Branson and Mehl,
Hindeodella corpulenta Branson and Mehl
Hindeodella germana Holmes
Hindeodella ibergensis Bischoff
Lambdagnathus macrodentata sp. nov.
Ligonodina obunca Rexroad
Ligonodina typa (Gunnell)
Lonchodina furnishi Rexroad
Lonchodina cf. *projecta* Ulrich and Bassler
Lonchodina? recurvata Bischoff
Mestognathus bipluti sp. nov.
Metalonchodina bidentata Branson and Mehl
Neoprioniodus inclinatus Hass
Neoprioniodus scitulus Branson and Mehl
Neoprioniodus singularis Hass
Neoprioniodus spathatus sp. nov.
Neoprioniodus varians Branson and Mehl
Ozarkodina delicatula Ellison
Roundya subacoda (Gunnell)
Spathognathodus sp.
Subbryantodus stipans Rexroad
Subbryantodus subaequalis sp. nov.
Synprioniodina forsenta Stauffer, 1940

The dominant genus is *Gnathodus* which provides over 50 per cent of the 2,000 specimens studied. The species of this genus show considerable infraspecific variation, and particularly so that species *Gnathodus commutatus* illustrated in Text-fig. 1. This figure, based on a study of 480 specimens, shows some of the main varieties of *Gnathodus commutatus* and some of the possible relationships between them. The most common form is *Gnathodus commutatus commutatus* (*a*) which has a smooth and almost circular cup. This simple form grades into *Gnathodus commutatus nodosus* (*b*) by the acquisition of nodes on the oral surface of the cup and into *Gnathodus commutatus homopunctatus* (*c*) by the acquisition of a platform and an oval cup. The single specimen of *Spathognathodus* sp. (*d*) may be an extreme variant related to *Gnathodus commutatus nodosus*.

The study of *Gnathodus commutatus* illustrates in like manner to the work of Scott and Collinson (1959), Rhodes (1952), Bischoff (1957), and others, that little reliance can be placed on ornamentation in the specific determination of Carboniferous conodonts. Similar variation within *Gnathodus commutatus* has been previously described from the P_2 zone of the German Carboniferous (Bischoff, 1957). There are, however, some differences in the variation exhibited by the German and British faunas, particularly in the numerical proportions of the variants, and which may be due to the fact that the present fauna represents the higher horizon E_1. Thus *Gnathodus commutatus nodosus*

described from the occurrence of a few specimens only from the German P_2 zone is at the higher E_1 horizon very abundant and variable, and may be an important index fossil.

Comparison of the Fauna with Faunas from America and Germany

The major content of conodont literature on the Carboniferous system has been concerned with the description of American faunas. Thus any comparison of the present fauna must of necessity be made with the American faunas. In so doing the problems of intercontinental correlation arise, and in particular that which concerns the American equivalent of the Lower Namurian and the Lower-Upper Carboniferous boundary.

A number of workers (Moore, 1937, Moore and Elias, 1937, Elias 1956, and Weller *et al.*, 1948), have considered this problem. Moore (1937), mainly on the evidence of goniatites, considered that the American Chester Series, of Upper Mississippian age, was equivalent to the Goniatites, *Eumorphoceras* and part or all of the *Homoceras* zones of Europe. He was uncertain of the position of the American equivalent of the European Lower-Upper Carboniferous boundary since *Eumorphoceras* seemed to appear earlier in America than in Europe. Elias (1956) arrived at the same conclusion regarding the age of the Chester Series but considered that the early forms of *Eumorphoceras* occurring in America were not typical of the genus as found in the *Eumorphoceras* zone of Europe. He therefore concluded that the Lower Namurian was equivalent to late Chester, and that the base of the Namurian occurred within the Chester Series.

Comparison of the American and present British conodont faunas proposes a correlation similar to that suggested by Elias. The majority of the genera represented in the British fauna are long-ranging forms, eleven range through the Carboniferous. The remaining genera are more restricted in range and are known to have the following ranges in the American Carboniferous.

Cavusgnathus	Upper Mississippian—Pennsylvanian
Geniculatus	Mississippian
Gnathodus	Mississippian—Upper Pennsylvanian
Lambdagnathus	Upper Mississippian
Subbryantodus	Mississippian

Mestognathus is known only from the Middle *Goniatites*—Stufe in Germany and is not known in America. Since the genera are long-ranging, species will be more important for detailed sub-division of this part of the Carboniferous sequence.

The conodont sequence of the type Chester Series in Illinois has recently been described (Rexroad, 1957, 1958), and comparison of

these forms with those described in the present paper shows eight genera and eleven species common to the two areas. The most important common forms are undoubtedly the genus *Lambdagnathus* and the species *Lonchodina furnishi*. The preliminary American studies suggest that these two forms are restricted to the Glen Dean Formation of the Chester Series.

The conodonts recorded by Elias (1956), and which occur in association with the goniatites *Cravenoceras* aff. *malhamense* and *Eumorphoceras bisulcatum*, are of little value for comparative purposes since the forms common with North Staffordshire range through most of the Mississippian.

In recent years a number of papers have described the German Lower Carboniferous conodont faunas and these show great similarities with the fauna described in this paper. Bischoff (1957), described faunal assemblages representative of horizons from the base of the Lower Carboniferous to the top of the *Goniatites*-Stufe (P-horizon). The fauna from the P-horizon contains thirteen genera, seventeen species and four sub-species which are now shown to be common to North Staffordshire and Germany. There are, however, important differences between the two faunas. The most important differences lie in the occurrence in North Staffordshire of the new species *Mestognathus bipluti*, and *Neoprioniodus spathatus*; in the abundance of *Gnathodus commutatus nodosus*; and in the presence of the genus *Lambdagnathus*, a form previously only described from the Glen Dean Formation of the Chester Series in Illinois. These differences are of importance particularly in view of the slightly younger age of the North Staffordshire conodonts as compared with the German assemblage.

Acknowledgements

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DEPARTMENT OF GEOLOGY,
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EXPLANATION OF PLATES

PLATE X

FIGS. 1, 2.—*Mestognathus bipluti* sp. nov. × 20, Oral views of holotype—1 (E1.E.1) and paratype—2 (E1.E1.2).

FIG. 3.—*Cavusgnathus unicornis* Youngquist and Miller. × 20, oral view (E1.E30).

FIG. 4.—*Gnathodus girtyi* Hass. × 20, oral view (E1.E15).

FIG. 5.—*Gnathodus bilineatus* (Roundy). × 20, oral view (E1.E1.4).

FIG. 6.—*Gnathodus commutatus* Branson and Mehl var. *commutatus* Bischoff. × 30, oral view (E1.E7).

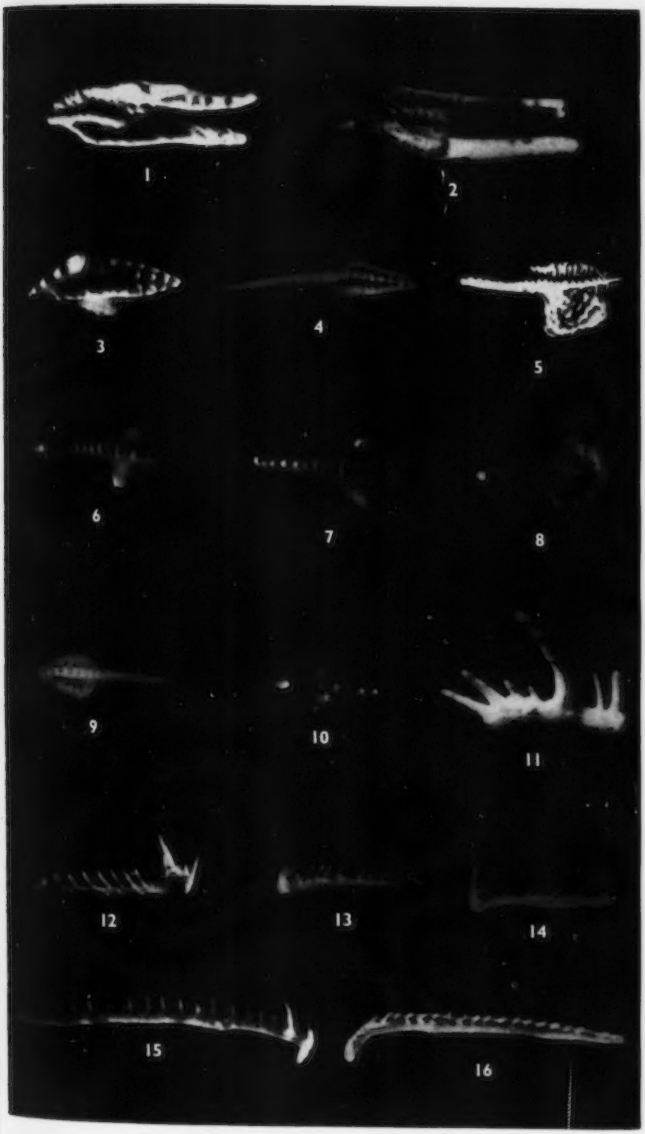
FIGS. 7, 8.—*Gnathodus commutatus* Branson and Mehl var. *nodosus* Bischoff. × 30, oral views 7(E1.E.8), 8(E1.E.9).
FIG. 9.—*Gnathodus commutatus* Branson and Mehl var. *homopunctatus* Bischoff. × 30, oral view (E1.B1).
FIG. 10.—*Spathognathodus* sp. × 30, oral view (E1.B10).
FIG. 11.—*Hindeodella corpulenta* Branson and Mehl. × 25, inner lateral view (E1.B6).
FIGS. 12, 13.—*Hindeodella germana* Holmes. × 25, inner lateral views, fig. 12 (E1.E27), fig. 13 (E1.X1).
FIG. 14.—*Hindeodella brevis* Branson and Mehl. × 25, oral view (E1.E2.1).
FIG. 15.—*Hindeodella ibergensis* Bischoff. × 25, inner lateral view (E1.E28).
FIG. 16.—*Angulodus walrathi* Hibbard. × 25, inner lateral view (E1.E31).

PLATE XI

FIG. 1.—*Neoprioniodus scitulus* Branson and Mehl. × 20, inner lateral view (E1.E15).
FIGS. 2, 4.—*Neoprioniodus spathatus* sp. nov. × 20, inner lateral views of holotype—4(E1.E22) and paratype—2(E1.E13).
FIG. 3.—*Neoprioniodus inclinatus* Hass. × 20, inner lateral view of young form (E1.E11).
FIG. 5.—*Neoprioniodus singularis* Hass. × 20, inner lateral view (E1.E12).
FIG. 6.—*Ligonodina typa* (Gunnell). × 25, posterior view (E1.E25).
FIG. 7.—*Neoprioniodus varians* Branson and Mehl. × 20, inner lateral view (E1.E22).
FIG. 8.—*Lonchodina furnishi* Rexroad. × 20, inner lateral view (E1.E16).
FIG. 9.—*Ligonodina obunca* Rexroad. × 20, inner lateral view (E1.E26).
FIG. 10.—*Lonchodina cf. projecta* Ulrich and Bassler. × 20, inner lateral view (E1.E18).
FIG. 11.—*Geniculatus claviger* Hass. × 15, inner lateral view (E1.E2.11).
FIG. 12.—*Lonchodina? recurvata* Bischoff. × 25, inner lateral view (E1.E17).
FIG. 13.—*Roundya subacoda* (Gunnell). × 20, posterior view (E1.E2.12).

PLATE XII

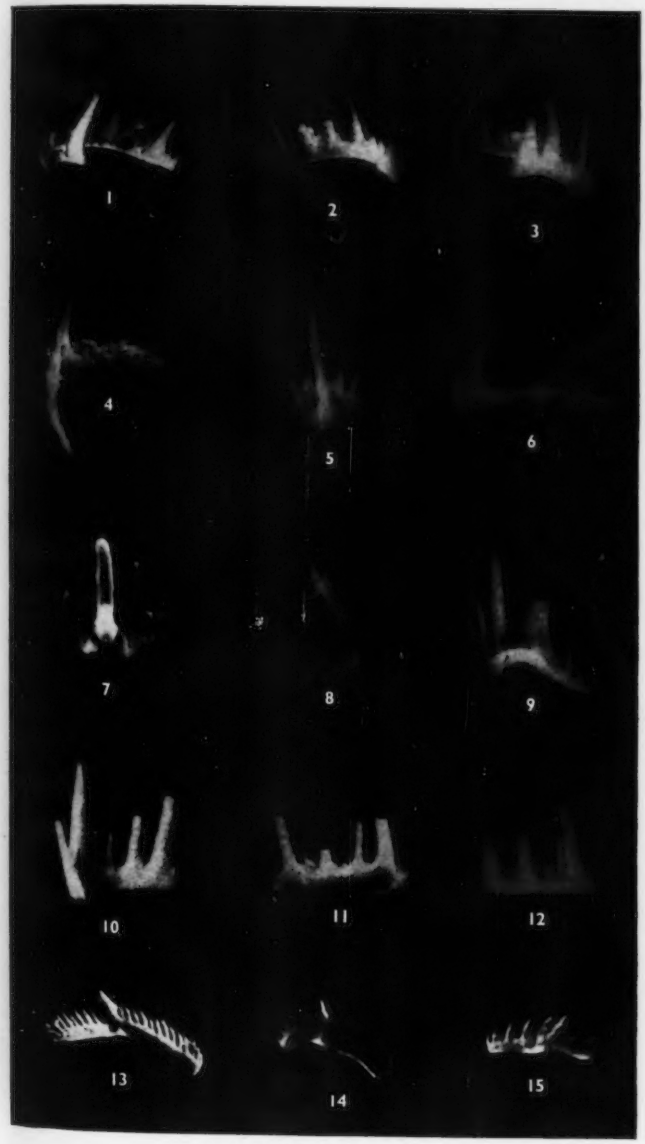
FIGS. 1–3.—*Lambdagnathus macrodentata* sp. nov. × 25, figs. 1 and 2, inner and outer lateral views of holotype (E1.E2.3); fig. 3, inner lateral view of paratype (E1.E2.4).
FIG. 4.—*Hibbardella fragilis* sp. nov. × 30, lateral view (E1.E2.11).
FIGS. 5, 6.—*Hibbardella pennata* sp. nov. × 30, fig. 5, posterior view of holotype (E1.E2.12); fig. 6, lateral view of paratype (E1.E2.15).
FIG. 7.—*Hibbardella milleri* Rexroad. × 30, posterior view (E1.E21).
FIG. 8.—*Synprioniodina forsenta* Stauffer. × 30, inner lateral view (E1.B3).
FIG. 9.—*Metalonchodina bidentata* Branson and Mehl. × 30, inner lateral view (E1.B4).
FIGS. 10–12.—*Hindeodella undata* Branson and Mehl. × 30, figs. 10, 11, lateral view of anterior end (E1.E2.6), (E1.E2.7), fig. 12, fragment of posterior bar (E1.E24).
FIG. 13.—*Ozarkodina delicatula* Ellison. × 20, lateral view (E1.E2.14).
FIG. 14.—*Subbryantodus stipans* Rexroad. × 20, lateral view (E1.E20).
FIG. 15.—*Subbryantodus subaequalis* sp. nov. × 30, lateral view (E1.E19).



NAMURIAN CONODONTS

facing p. 224.

NAMURIAN CONODONTS



NAMURIAN CONODONTS

Lepidostrobophyllum fimbriatum (Kidston 1883) from the Drybrook Sandstone (Lower Carboniferous)

By K. C. Allen

(PLATE XIII).

Abstract

Lepidostrobus fimbriatus Kidston is redescribed as *Lepidostrobophyllum fimbriatum* (Kidston) from a new locality, in the Drybrook Sandstone of the Forest of Dean. The sporophyll consists of two parts bent at right angles, forming a heel of hard tissue. The sporangium is attached to a small area by the midrib, and is only partly occupied by the four spiney megaspores. A ligule is present.

Introduction

Lepidostrobophyllum fimbriatum, present as detached sporophylls, is the most common species in the flora of the Drybrook Sandstone from a quarry at Puddlebrook (Nat. Grid Ref. 646.184) in the Forest of Dean. The flora was collected from the lower of the two irregular shale bands which occur within the dominant sandstones and conglomerates of the quarry. Only one species has previously been recorded from the Drybrook Sandstone. Wethered (1883, p. 215) recorded *Lepidodendron* sp., allied to *L. griffithii* (Brogniart), and Sibly (1912, p. 420) stated that no other determinable plant fossils were present. Recent collecting has shown the presence of several species in addition to the one now described, including *Lepidostrobophyllum* sp., *Halonia* sp., and several fern-like leaves.

Preservation is poor and little plant substance now remains, suggesting that the carbonaceous material has undergone slow oxidation in a damp state, although the specimens may once have been good compressions. In many specimens, virtually no organic matter remains. Although traces of cuticle are present, they proved of little use. Balsam transfers gave no new information.

I should like to express my thanks to Professor T. M. Harris for his encouragement and criticism, to Dr. I. Strachan for the loan of material, and to Dr. W. S. Lacey for unpublished information.

Figured specimens (other than Text-figs. 1, 3, and 5) have been presented to the British Museum (Natural History). The originals of Text-figs. 1, 3, and 5, are in the Geology Department at Birmingham University and the remainder of the collection is in the Museums of the Geology Departments of Reading and Birmingham Universities.

Lepidostrobophyllum fimbriatum (Kidston, 1883)

(Plate XIII, figs. 1–5; Text-figs. 1–5)

1883. *Lepidostrobus fimbriatus* Kidston, p. 543, pl. 31, figs. 2, 3, 4. (Type and other specimens from S. Scotland and Northumberland).
1905. *Lepidostrobus fimbriatus* Kidston, p. 823–5. (Recorded from Cumberland.
1952. *Lepidostrobus fimbriatus* Lacey, p. 20. (Recorded from N. Wales).
1952a. *Lepidostrobus fimbriatus* Lacey, p. 375. (Recorded from N. Wales).

1
2
3
4
5

Emended diagnosis

Isolated sporophylls (megasporophyll alone recognized). Length of sporophyll about 24 mm. (range 17–28 mm.) composed of a vertical limb 7–12 mm. long, and a horizontal limb 10–16 mm. long. Sporophyll about 14 mm. wide (range 12–18 mm.) where the two limbs join, upper limb tapering to an acute point, lower tapering to a truncate base 5 mm. wide. Vertical limb composed of a thickened midrib and thickened marginal ribs, intervening tissue delicate, margins bearing up to twenty short outward pointing processes about 2·5 mm. long (range 1·2–3 mm.). Marginal ribs of horizontal limb broad without processes; at angle between limbs, the thickened tissue of midrib and margins extends laterally to unite as a transversely thickened bar. Sporangium rounded, 5–8 mm. wide, borne on midrib of horizontal limb, overlapping a short, broad ligule; megaspores four, body of spore rounded 1·2–1·9 mm. wide, with slender tapering spines up to 0·3 mm. long. Spore without equatorial flanges or strongly developed apical processes.

Description

The sporophylls are composed of two limbs, set almost at right angles to one another, one being horizontal, the other almost vertical. When compressed, the original attitude of the two limbs is not always evident. Either there is no indication from the compressions that they were not both in the same plane (Pl. XIII, fig. 2) or the sporophylls are broken at the junction between the two limbs (Pl. XIII, fig. 5). However, in specimens compressed on their sides, the angle between the limbs can be clearly seen (Text-figs. 1, 2).

The shape of the vertical limb is seen in its typical form in Plate XIII, fig. 5. The margin, which bears up to twenty fimbriations, is thickened; so too is the midrib. The area between is occupied by a delicate tissue, which is often invisible, but in the best specimens it is seen to consist of small cells. At the base of the vertical limb, the thickened tissues expand into a broad band, which forms a kind of heel.

The horizontal limb (Text-fig. 3) shows a much wider outer zone of thickened tissue, which is not fimbriate. The midrib is seen in Text-fig. 1, but in the specimens of the other figures, it is either concealed by the sporangium or broken at the angle of the sporophyll, and in Plate XIII, fig. 1 it is split into a fine central strand

TEXT-FIG. 1, 2.—Sporophylls showing the angle between the two limbs (×3). Fig. 2, B.M.N.H., V42329.

TEXT-FIG. 3.—Underside of the sporophyll showing the midrib of the horizontal limb (×3).

TEXT-FIG. 4.—A group of megaspores (×10), showing the general shape and spines. B.M.N.H., V42333.

TEXT-FIG. 5.—Specimen showing the difference in width of the outer thickened areas of both limbs (×3).

which might be vascular, and two lateral strands which might be fibrous. There is presumably delicate tissue between the midrib and the thickened margins, but nothing is preserved, and most of this area is covered by the sporangium.

The position, size, and shape of the sporangium is shown in Plate XIII, fig. 5, where it slightly overlaps onto the vertical limb. When this overlapping front part in Plate XIII, fig. 4 was removed, a broadly rectangular body was visible. It is seen also in Plate XIII, fig. 2. This body is above the midrib, and separated from it by the thin layer of matrix. It is presumed to be the ligule.

Each well preserved sporangium has four megaspores, which occupy barely half the sporangium (Plate XIII, fig. 5). The megaspores are towards the outer end of the sporangium. The substance of the megaspore wall is ill preserved and even the best yielded very poor preparations of cutinized fragments. All that can be seen, is that the spores are 1·2–1·9 mm. in diameter, and they have sharp spines up to 0·3 mm. long. It is certain that the spores have no equatorial flanges nor large apical prominences, and even triradiate marks were not visible. All that could be seen in the best specimen is indicated by Text-fig. 4.

Certain sporophylls (not figured) differ in that the sporangia are of lighter colour and show no trace of megaspores. Fragments of these sporangia were gently macerated. They yielded a spongy material different from the cutinized remains of the megasporangia, but nothing could be recognized as microspores. However, it remains possible that the spongy matter consists of a mass of microspores, and if so the microsporophylls are of exactly the same shape and appearance as the megasporophylls.

Discussion

It is unfortunate that the megaspores are poorly preserved and cannot be precisely identified with any dispersed spores. However, the absence of an apical process or an equatorial zone and the presence of tapering spines, points to a fairly small group of spore species. Of these it is most like *Triletes echinoides* Chaloner. However, *T. echinoides* found only in Indiana (Chaloner, 1954, p. 29) is sometimes at least, larger (known range 2·0–2·6 mm. as against 1·2–1·9 mm.), and the spines are much longer (0·9 mm. as against 0·3 mm.). It is to be hoped that rather better specimens of *L. fimbriatum* will give good spores.

Three other spores known of similar size and with similar spines are *T. horridus* Zerndt, *T. subpilosus* Ibrahim and *T. crassiaculeatus* Zerndt. These have well-developed apical appendages which seem to be absent here.

Distribution and Age Range

Lepidostrobophyllum fimbriatum is recorded in Britain only from a

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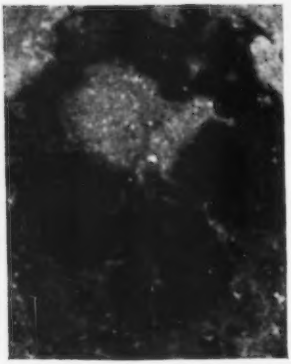

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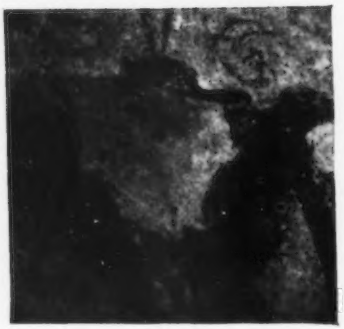

LEPIDOSTROBOPHYLLUM FIMBRIATUM (KIDSTON)

facing p. 23

few districts. Kidston (1883) described the species from the Oil-Shale and Cementstone group of Scotland and Northumberland. In 1905 he recorded it from the Scremerstone Beds in Northumberland and from the Cementstone group in Cumberland. In these localities the exact ages are problematical, but all are of Lower Carboniferous (C_1–D) age. Lacey (1952, 1952a, 1955) recorded the species, noting also tetrads of megaspores similar to *T. echinoides*, from the Lower Brown Limestone of North Wales which are of S_2D_1, or early D_1 age. The present specimens from Drybrook are of S_2 age. *L. fimbriatum* therefore appears to be confined to the Lower Carboniferous (C_1–D), the best zoned specimens being from the S_2D_1 zones.

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EXPLANATION OF PLATE XIII

Fig. 1.—*Lepidostrobophyllum fimbriatum* (Kidston), (×4) showing the broad band of thickened tissue at the angle of the limbs. B.M.N.H., V42330.

Fig. 2.—A specimen (×4) showing general shape. B.M.N.H., V42331.

Fig. 3.—A group of three megaspores and an impression of the fourth (×10). Spines can be seen on the lower right megaspore. B.M.N.H., V42333.

Fig. 4.—Horizontal limb (×4) with the sporangium removed, showing the rectangular ligule. B.M.N.H., V42332.

Fig. 5.—Sporophyll (×3) showing the four spiny megaspores in the megasporangium. The megasporangium overlaps onto the vertical limb. B.M.N.H., V42333.

On the species *Orbignyella fibrosa* (Lonsdale)

By David E. Owen

PLATE XIV

Abstract

The species *Alveolites fibrosa* Lonsdale, 1839, described also as *Stenopora fibrosa* var. *β Regularis* McCoy, 1851, and referred to as *Stenopora fibrosa* var. *incrustans* McCoy, was recognized originally by Lonsdale (1839) as different from *Calamopora fibrosa* Goldfuss, 1826. It is described here as *Orbignyella fibrosa* (Lonsdale) and shown to occur typically in beds of Leintwardine age.

Introduction

In collecting bryozoa from the Ludlovian rocks of the Ludlow-Wigmore road, section by section, I found an encrusting form in beds of Lower Leintwardine age. In sectioning a specimen labelled *Stenopora fibrosa* var *incrustans* McCoy from the British Museum collections (No. PD. 3520), I realized that it was the same species. It was also clearly *Alveolites fibrosa*, Lonsdale, 1839. Finally, in collecting in a muddy limestone above the massive limestones of Perton Quarry on the north western edge of the Woolhope Dome I found a few normal, encrusting forms, one on a small gastropod shell, and a very large number of stout ramose forms of the same species.

Systematic Description

Order : Trepostomata, Ulrich, 1882.

Family : Monticuliporidae, Nicholson, 1879, emended Ulrich 1890.

Genus : *Orbignyella*, Ulrich and Bassler, 1904.

This genus was defined to include those Monticuliporid species which differ from the genus *Monticulipora* in the following three features. The zooecial walls are lacking in the irregular granulose structure of that genus, the cystiphragms are represented by structures more like curved diaphragms and the acanthopores are well developed and sharply defined.

Most of the species so far described occur in Ordovician rocks, though the lamellate species *Orbignyella expansa* (Ringueberg) and *Orbignyella magnopora* Bassler have been described by Bassler from the Rochester Shales (1906). Neither could be confused with the species about to be described.

Orbignyella fibrosa (Lonsdale)

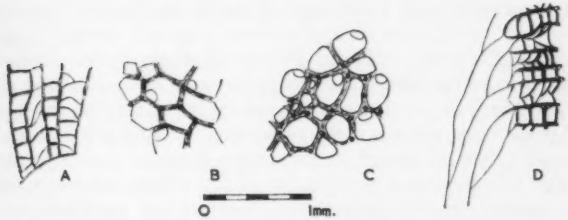
Plate XIV, figs. A, B, E, and F. Text-fig. 1 A and B

1839. *Alveolites, ? fibrosa* Lonsdale, 681, pl. 15, figs. 1 and 1a, (nec. *Calamopora fibrosa* Goldfuss, 1826, 82, pl. 28, figs. 3a, 3b, 4a, and 4b. nec. *Favosites fibrosa* Lonsdale, 1839, 683, pl. 15 bis. figs. 6, 6a–f. nec. *Favosites fibrosa* Phillips, 1841, 17, pl. 9, figs. 25a–c. nec. *Aveolites fibrosus* d'Orbigny, 1850, 1, 108. nec. *Favosites fibrosa* E. and H., 1853, 4, 217, pl. 48, figs. 3, 3a, and 3b, and 1854, 5, 261, pl. 61, figs. 5 and 5a.)

1850. *Alveolites subfibrosus d'Orbigny*, 1, 49.

1851. *Stenopora fibrosa* var *β Regularis* McCoy, 25.

Diagnosis.—Zoarium usually encrusting, sometimes covering small gastropod shells to a depth of up to 3 mm., sometimes coating thinly other shells or grains, and sometimes occurring as an incrustation on some material which has disappeared completely. Zooecia straight, fairly thick-walled 0·15–0·2 mm. diameter, with straight diaphragms in some tubes occurring at about tube width intervals, and curved diaphragms or cystiphragms, running down from one or both sides of the tube, in adjoining tubes. Sometimes both straight and curved diaphragms occur in the same zooecial tube. True mesopores are absent, but zooecia of much smaller diameter around 0·08–0·1 mm.



TEXT-FIG. 1 A–D.—Vertical and tangential sections of *Orbignyella fibrosa* (Lonsdale) (A and B), and *Orbignyella fibrosa* (Lonsdale) forma *ramosa* (C and D).

which are not otherwise markedly different, are fairly abundant. Acanthopores small, clearly defined, occur in the mature region and are seen in some tangential sections to be circular in outline with a minute hollow centre. Apertures oval to polygonal. As tangential sections often cut the curved diaphragms, these appear as small arcs on one side of a zooecium.

Widespread in areas where Ludlovian rocks occur, this species does not seem to be recorded in rocks below the Lower Leintwardine series or to extend as far up in the succession as the Lower Whitcliffe beds.

Remarks.—When Lonsdale was describing fossils for Murchison's *The Silurian System* (1839), he named, wrongly as it subsequently turned out, a group of hemispherical bryozoa *Favosites fibrosa* (Goldfuss). Some of these forms have been re-examined, and those from the Upper Ordovician shown by Spjeldnaes (1957), to belong to the species *Batostoma murchisoni* nom. nov. Bryozoa from the Wenlock Limestone described by Nicholson (1884) as *Monotrypa crenulata* are undoubtedly the same as Lonsdale's Wenlock Limestone forms (Owen, 1960).

At the same time, Lonsdale had a number of specimens encrusting

gastropod shells from the Upper Ludlovian and " Aymestry Limestone " which he felt were a new species of the genus *Alveolites*. He describes how he first thought them to belong to the species *Favosites fibrosa* (Goldfuss), but " The beautiful specimen plate 15, figure 1, was afterwards added to the collection : and from the regularity of the surface of each layer I have been induced to remove the coral into the genus *Alveolites* ". He retained the trivial name *fibrosa*. Therefore, despite the confusion of different genera bearing the same trivial name *fibrosa*, as this species is clearly figured, there is no option but to retain Lonsdale's name for it.

In 1850, d'Orbigny recorded *Alveolites subfibrosus* d'Orbigny (p. 49) as the same species as Londsdale had figured as *Alveolites fibrosa* and noted that it came from the Ludlow-rock, England. He also recorded from the Devonian, *Alveolites fibrosus* d'Orbigny (p.108) which he believed to be the same species as *Calamopora fibrosa* Goldfuss. He neither described nor figured these species but referred to the figures and descriptions of other authors including Lonsdale and Goldfuss.

In 1851, McCoy placed all the species *fibrosa* in Lonsdale's genus *Stenopora* (once more wrongly, as it has subsequently appeared). The species *Favosites fibrosa* (Goldfuss) became *Stenopora fibrosa* (Goldfuss) ; *Alveolites fibrosa* Lonsdale became *Stenopora fibrosa* var. β *Regularis* McCoy. Specimens in the British Museum collections (PD. 3515–20) encrusting gastropods and clearly belonging to this variety are labelled *Stenopora fibrosa* var. *incrustans* McCoy.

In 1853 and 1854, Edwards and Haime described the " corals " of the Devonian and Silurian rocks respectively. Everything in the *fibrosa* group from Ordovician to Devonian is described as *Favosites fibrosa* (Goldfuss).

In 1884, Nicholson, the first palaeontologist to describe corals and bryozoan species on internal structure as revealed in thin sections viewed under the microscope, published the species *Monotrypa crenulata* Nicholson, which he believed to be the same species as the Wenlock Limestone examples of Lonsdale's *Favosites fibrosa*. He did not retain the name as he knew that Lonsdale's species was different from Goldfuss' species. Unfortunately, he did not know *Alveolites fibrosa* Lonsdale.

This species is clearly a monticuliporid and it fits perfectly the description of the Genus *Orbignyella* with its sturdy structure and equal development of straight and curved diaphragms, and small but well-marked acanthopores.

Careful collecting in the Ludlow sections, particularly along the Ludlow-Wigmore road, shows that *Orbignyella fibrosa* (Lonsdale) does not appear in the Eldon or Bringewood beds. It is a common species, however, in the Lower Leintwardine beds, forming 20 per cent of the

bryozoa determined. Its range does not extend up to the Whitcliffe beds. It may thus prove to be a useful indicator of beds of Leintwardine age.

Orbignyella fibrosa (Lonsdale) forma *ramosa*.
Plate XIV, figs. C, D, G and H and Text-fig. 1 C and D.

Diagnosis.—Zoarium stout, cylindrical, ramose, diameter 2–7 mm. and length up to several centimeters. Zooecia in central, immature region, run nearly parallel with the axis and are very thin-walled. At 0·5–2 mm. from the surface they turn sharply and run straight to reach the surface at right angles to the axis. In this mature region the walls are thick and the zooecia identical to those in the encrusting forms.

Remarks.—These ramose forms occur side by side with normal encrusting forms. I have found them only in a bed of muddy limestone lying on top of the massive limestones of Perton Quarry east of Hereford on the north western tip of the Woolhope Dome. This bed is quite thin, but is immensely rich in specimens of this single species. These are described as a form and not a variety of *Orbignyella fibrosa* (Lonsdale) as they occur with the more typical forms and were perhaps induced by specially favourable conditions and an absence of shells or other encrustable objects.

Acknowledgments

The author would like to thank Dr. H. Dighton Thomas for his advice on the retention of Lonsdale's species, Dr. E. V. Tucker for drawing his attention to the Perton locality and Mr. Harry Spencer for photographing the solid specimens.

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EXPLANATION OF PLATE XIV

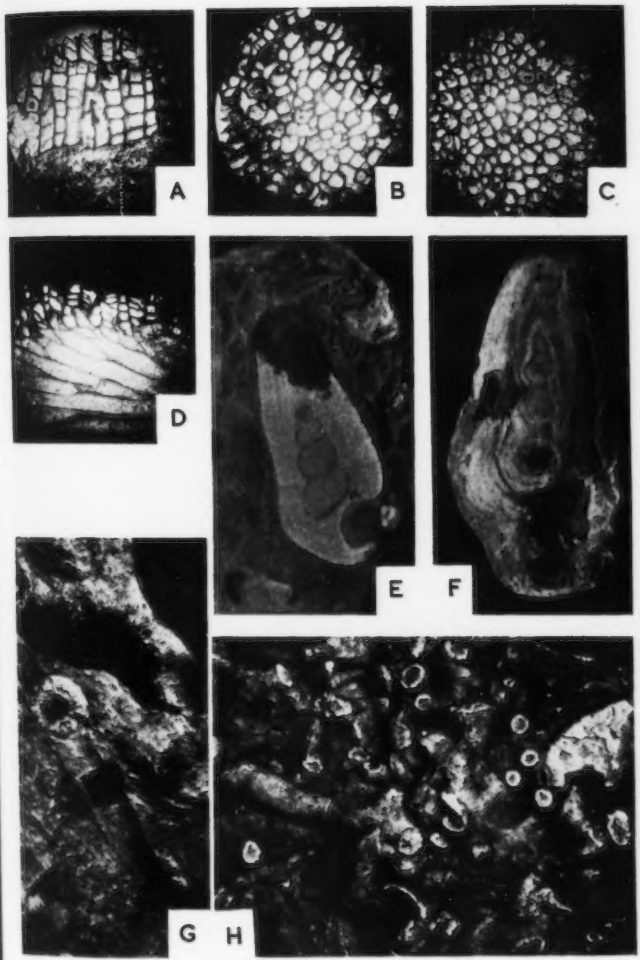
A and B.—*Orbignyella fibrosa* (Lonsdale), vertical section showing straight and curved diaphragms, tangential section showing curved diaphragms in section. B.M.N.H., PD. 3520. X 13.

C and D.—*O. fibrosa* (Lonsdale) forma *ramosa*, tangential section showing acanthopores and curved diaphragms in section, vertical section showing thin-walled immature region and straight and curved diaphragms in mature region. Manchester Mus., LL. 2537 and 2538. X 13.

E.—*O. fibrosa* (Lonsdale), polished specimen encrusting gastropod. B.M.N.H., PD. 3515. X 1½.

F.—*O. fibrosa* (Lonsdale), encrusting gastropod. B.M.N.H. PD. 3516. X 2.

G and H.—*O. fibrosa* (Lonsdale) forma *ramosa*, ramose forms from Perton Quarry near Woolhope. Manchester Mus., LL. 2535 and 2536. X 1.



ORBIGNYELLA FIBROSA (LONSDALE)

facing p. 234.

Pre-Cambrian Perthosites in Nyasaland

By S. W. Morel

Abstract

Attention is drawn to the widespread occurrence of perthosite bodies in the Basement Complex of southern Nyasaland, and a description of these rocks is given.

It is suggested that these rocks represent a late stage in the differentiation of basaltic or andesitic magma which produced basic intrusives and volcanics and ultrabasic complexes in the late Pre-Cambrian. These were subsequently metamorphosed at a high grade. The nepheline-syenite gneisses of Port Herald and Tambani are believed to be related to the same series.

The remarkable similarities between the late Pre-Cambrian igneous history and that of the Karroo to Chilwa Series period is commented upon, and the occurrence of similar " ring structuress " in both series is described.

Introduction

The rock *perthosite* was so named by Phemister (1926) to describe an almost monomineralic saturated syenite composed of crystals of antiperthitic feldspar. At the type locality, Loch Ailsh, this rock intrudes an earlier series of nordmarkites, pulaskites, shonkinites, biotite-pyroxenites and hornblendites. The accessory minerals of the perthosite include aegirine-augite, sphene, zircon and opaque oxides.

Perthosites are not common rocks and in view of the many carbonatite vents and associated alkaline igneous intrusions that have been described from Africa in recent years, it may be of interest to draw attention to the widespread development of alkaline intrusives in the Basement Complex of southern Nyasaland at the close of Pre-cambrian times.

There are two series of alkaline intrusives in southern Nyasaland, a late Pre-Cambrian one which is typically associated with basic and ultrabasic intrusives, and a post Karroo series of nordmarkites, perthosites and nepheline-syenites related to the Chilwa volcanic episode (Text-fig. 1). The latter series form the great mountain masses of Zomba and Mlanje and have been described by Dixey (1937, 1955), Garson (1958), Bloomfield (1952), and Stringer, Holt, and Groves (1956). A characteristic feature of the Chilwa series rocks is the so-called " ring structure ". This paper is concerned with the first of these two series of rocks.

The Perthosites

The Pre-Cambrian perthosites are generally foliated or lineated rocks composed principally of flesh-pink perthite crystals up to 1 cm. in diameter. A small percentage of mafic minerals is present, consisting mainly of hornblende and biotite, which increase in amount towards

the borders of the intrusions, where they give the rock an irregular foliation. The central parts of the larger bodies are often unfoliated and are more resistant to forces of erosion. It is these unfoliated portions that build the mountain masses such as Chiradzulu, Michiru,

Text-fig. 1. Geological Map of Southern Nyasaland.

and Ndirande which rise out of the mid-Miocene Shire Highlands erosion surface.

The perthosites merge indistinctly into the surrounding banded granulites and amphibolites. A traverse across the boundary of the perthosites reveals a gradual decrease in the size and number of the perthite phenocrysts whilst a foliation becomes more and more distinct. It becomes evident that mafic schlieren-like bands in the perthosites represent basic members of the surrounding charnockitic granulites or bands of amphibolite which have undergone alkali

metasomatism during the emplacement of the perthosite (Morel, 1958, p. 49).

In thin section the perthite is seen to be the type that Michot (1951) has called "mesoperthite". The perthite crystals are invariably anhedral with crenulated borders interlocking one with the other. They vary considerably in grain size but average 1–2 mm. in diameter. String perthite is the commonest type but film, vein, and patch perthite are also generally seen. The twinning of the perthite is usually too fine to determine the individual components with a low-power objective, except where it contains isolated patches of albite in the patch perthite. Small twinned anhedral grains of oligoclase are common in the film perthite and this mineral also forms anhedral grains between the perthite phenocrysts. These grains may be sufficiently common in some examples to separate the individual perthite phenocrysts from each other. Myrmekite is rare but in some examples it is well developed with twinned oligoclase lobing into the film perthite and enclosing typical dendritic blebs of quartz.

A little quartz is occasionally observed between the perthite crystals. In some localities this increases considerably in amount and the rocks grade into quartz-syenites consisting entirely of perthite, quartz, and oligoclase.

The perthosites vary considerably in their content of mafic minerals. Of these biotite is the commonest and forms flakes which are pleochroic from pale yellow to brown and sometimes altered to green chlorite. It tends to form around grains of iron ore. Green hornblende forms discrete anhedral crystals and also spongy intergrowths surrounding partly altered green augite. The hornblende is pleochroic from pale olive-green to deep green. The augite is pale green in colour and is invariably partly altered to hornblende. It appears to be a similar variety to that found in the charnockitic granulites which form the country rocks. A few crystals of strongly pleochroic hypersthene with numerous schiller inclusions and embayed boundaries have been seen in specimens collected near to the margins of the intrusions. Magnetite and ilmenite are frequently present in these rocks and in some cases they vein fractures in the perthite crystals. Apatite, sphene, and minute zircons are present in accessory amounts as rounded grains and allanite has also been observed. Rare flakes of muscovite are occasionally seen between the perthite phenocrysts.

The perthosite bodies vary very greatly in size and shape. In the Ncheu area (Morel, 1957) they form sill-like bodies up to 300 feet thick striking parallel to the strike of the surrounding ortho- and paragneisses, and are repeated at intervals of a few hundred yards. In the Shire Highlands they cover large areas and appear to be of batholithic proportions, whilst at Chingali (Morel, 1951) and Mlindi (Ashley, 1951)

the structures are definitely of " ring structure " type with a central core of biotite-pyroxenite surrounded by a partial ring of foliated perthosite-gneiss. These would appear to represent the roots of volcanic complexes of similar type to the later post-Karroo " ring complexes ".

Nepheline-syenite-gneisses occur in the Basement Complex in the Port Herald hills (Bloomfield, 1958) and at Tambani (Cooper, 1946, 1952). The latter occurrence has provided zircons and monazite which have given an approximate age determination of 550,000,000 years (Tilley, 1953). These rocks are believed to be of similar age to the perthosite-gneisses and this is also thought to be the age of the Mozambique metamorphism which has produced the gneissic rocks of the Basement Complex.

Intrusions of ultrabasic rocks ranging from biotite, biotite-pyroxenite, websterite, and olivine-biotite-pyroxenite are common in the gneisses of the Basement Complex. They have suffered from the Mozambique metamorphism and have been subject to shearing, faulting, and alkali metasomatism (Morel, 1955, 1958). Where ultrabasic bodies have been engulfed by later intrusions of perthosite they appear as biotite-rich masses and schlieren within the perthosite. Where perthosite lies adjacent to websterite as at Mpemba mountain, the leucocratic perthosite passes without discordance into the mafic pyroxenite, the passage being marked by a gradual development of perthite augen in the pyroxenite until only a few bands rich in biotite remain to show that the perthosite has replaced the pyroxenite.

Although the margins of the perthosites are foliated there is some doubt as to whether this is due entirely to replacement of pre-existing foliated rocks or whether it is the result of fluxioning in incompletely consolidated perthosite magma at the same time as the surrounding gneisses were undergoing metamorphism at granulite and upper amphibolite facies grades.

The Lirangwe Complex

Some 15 miles to the N.N.W. of Blantyre three hills rise from an almost flat plain which is gently inclined north-westwards towards the Shire river. These hills were examined by Savage (1934) who reported that they were formed of perthitic syenite, norite, and pyroxenite. This area has recently been examined by Bloomfield, who believes that a wave of potash metasomatism passing through the basic and ultrabasic rocks during the waning stages of the regional metamorphism is responsible for the formation of these perthosites (1957, p. 20).

The largest of the three hills, Little Michiru (2,610 feet), is formed of a dark-green websterite which extends southwards across the Lirangwe river as a tongue 1 mile in width and 4 miles in length. At its southern

extremity it merges into biotite-pyroxenite and biotite-norite whilst to east and west it is flanked by perthitic norite (Morel, 1958).

Exposures in the bed of the Lirangwe river show that the websterite is intimately mingled with the perthitic morite. The latter occurs as veins, schlieren and segregations within the websterite as well as surrounding it. Shearing and intense granulation are evident in both rock types, and they have both undergone the regional metamorphism that has affected the country rocks.

The *websterite* consists essentially of an agglomeration of anhedral schillerized grains of hypersthene and augite which are often bent, fractured, and granulated. The interstices between the pyroxenes are occupied by pale-green to brown hornblende, apatite, biotite, and andesine or labradorite.

The *biotite-pyroxenites* are similar in mineralogy and texture to the websterite but contain rather more biotite and plagioclase. By a further increase in the amount of plagioclase these rocks grade into a dark-grey, fine-grained *biotite-norite*. This contains subhedral to euhedral crystals of diallage and schillerized hypersthene which are frequently twinned. Many of these crystals show an incipient alteration to green hornblende. The feldspar is usually andesine or labradorite although small granulitic crystals of oligoclase are present in some cases. The andesine and labradorite forms large twinned crystals up to 3 mms. in length which appear to have crystallized interstitially to the pyroxenes. The plagioclase twin-planes are bent and the rock has undergone cataclasis.

The *perthitic norite* is a grey rock with a high colour index. Its grain size varies from fine to medium depending on the amount of perthite present as phenocrysts although coarse-grained schistose types with perthite augen 1 cm. in length have been observed. A faint foliation is generally visible in the rock and is formed by regularly disposed flakes of biotite.

The perthitic norite exhibits intrusive junctions against the country rock which here consists of charnockitic granulites and biotite-gneisses. The latter contain augen of microcline, and the gneisses are feldspathized and migmatic with angular mafic xenoliths in a feldspathic matrix.

The perthitic norites vary considerably in their content of pyroxenes and perthites. The pyroxenes are similar to those found in the pyroxenite and biotite-norite with strongly pleochroic anhedral grains of hypersthene containing very numerous small brown platy schiller inclusions. The hypersthenes attain a size of 1 to 2 mm. but their margins are indentated and contain numerous cavities infilled with plagioclase, biotite, iron ore, and apatite. The diallage crystals are very similar in appearance to the hypersthenes and are pale green in colour. Secondary green hornblende and biotite are common in the pyroxene

cleavages. The amount of pyroxene present varies from about a tenth to a third of the rock.

The mafic minerals are surrounded by a fine-grained mosaic of oligoclase of granulitic habit. Small angular crystals of pyroxene occur between the oligoclase crystals and have the appearance of fragments dislodged from larger crystals by attrition during movements. Similar textures were observed in the pyroxenites and biotite-pyroxenites.

Biotite is usually present and imparts a weak foliation to the rock. It is strongly pleochroic from pale straw-yellow to deep brown and occurs in flakes up to 1 mm. in length, which commonly enclose grains of iron ore.

Tabular crystals of apatite up to 1 mm. in length are common accessories and ilmenite, magnetite, and iron sulphides form anhedral grains in many of the specimens examined.

The amount of oligoclase present varies with the amount of perthite. In the perthite-free types the plagioclase forms a fine-grained mosaic in which the pyroxenes are set. This mosaic has a distinctly granulitic appearance and consists mainly of acid-oligoclase, much of which is not visibly twinned. Rare phenocrysts of plagioclase are also present, attaining lengths of half a centimetre. They are invariably bent and fractured and have granulated boundaries. These phenocrysts are usually labradorites but reverse-oscillatory zoned bytownites have also been observed. Many of the oligoclase grains enclose minute rounded zircons. A little myrmekite is occasionally seen between the oligoclase grains in the groundmass which suggests that some untwinned orthoclase may be present. These oligoclase grains which form the groundmass seldom exceed 0·5 mm. in diameter and average about 0·2 mm.

The proportions of oligoclase and perthite vary considerably in the perthitic norite. In some parts the rock is similar to the perthite-free type described in the previous paragraph, but merely contains rare wisps of antiperthite in some of the groundmass oligoclases. In other cases the amount of perthite varies from a few crystals per thin section to rocks containing large numbers of perthite crystals which are only separated from one another by a narrow zone of granulitic oligoclase. A final stage is reached where little or no oligoclase is present and the feldspar consists almost entirely of anhedral perthite crystals, the crenulated boundaries of which interdigitate with each other. Apart from a slightly higher content of mafic minerals this rock is similar to the foliated perthosites found in the Basement Complex.

As the proportion of perthite in the rock increases, it forms veins around the margins of the oligoclase crystals and along cracks in them. In some cases this has progressed so far that only isolated remnants of oligoclase remain in the centres of the new perthite crystals. In the

smaller perthites braid and string perthite are the commonest modes. The larger grains are mesoperthites, exceed 1 cm. in size and are twinned on the Carlsbad law. In some of them the lamellae make an angle of 76° with the composition plane of the twins which suggests that they may be anorthoclase, but in many other cases the extinction angle approaches 10° as in microcline-perthite.

Some of the perthite crystals are formed of contiguous crystals of oligoclase, the perthitic portions of which coalesce to form the anhedral outlines of a perthite crystal, the central portion remaining as several separate oligoclase grains.

The lobes and embayments of the edges of the perthite crystals approximate to the grain size of the oligoclase groundmass mosaic. Some of the oligoclase grains adjacent to the perthite crystals are in optical continuity with the plagioclase component of the perthite, and in some of these albite twinning extends into the latter.

Small patches of microcline have been observed in some of the perthite crystals but these are of very rare occurrence.

The perthite crystals and the oligoclase grains immediately adjacent to them contain minute dark inclusions. In the perthites these are regularly disposed but they lie at random in the oligoclase. None of the perthite crystals observed in thin section appears to have suffered deformation and thus they appear to have crystallized after directional forces had ceased. In the well-foliated portions of the perthitic norite the biotite flakes have a regular and parallel disposition to each other. However, in some cases small flakes of biotite are enclosed in the perthites and are orientated parallel to the lines of dark inclusions, suggesting that their orientation was controlled by the porphyroblastic development of the perthite.

Dykes of biotite-norite are found in the perthitic norite. A faint foliation passes through both dykes and host, and porphyroblasts of perthite have grown in both the dykes and the perthitic norite (Morel, 1958, Plate IV).

In other localities at Lirangwe, Mpemba, and Ncheu intrusive dykes of perthosite have been emplaced into pyroxenites, olivine-biotite-pyroxenites, and Basement Complex gneisses.

The Origin of the Perthosites

The perthitic norite of the Lirangwe complex is clearly a magmatic intrusion of noritic type that has been metamorphosed following its consolidation. At a later stage porphyroblasts of perthite grew in this rock.

Michot (loc. cit.) believes that mesoperthite is formed by exsolution from an original homogenous phase. Heier (1955) believes that the mesoperthites of the Ørsdalen area are due to replacement. The

Ørsdalen mesoperthites appear to be similar to the Nyasaland perthites and like them are found in rocks which have suffered granulite facies metamorphism.

There would appear to be at least two possible origins for the perthites of the Lirangwe perthitic norite. First, the perthitic norite may represent an original syenogabbro which has suffered metamorphism resulting in the exsolution of potash feldspar from original plagioclase. Secondly, the presence of great volumes of perthosite in the Basement Complex shows that an ample source of alkalis was available to produce perthites by replacement. Perhaps the fact that the perthite of the Lirangwe rock is an antiperthite and possibly anorthoclase, would suggest that the first alternative is the correct one.

The total volume of the perthosite bodies in the Basement Complex greatly exceeds that of all the basic and ultrabasic intrusives with which they are associated. At first sight this would appear to rule out the possibility that they have arisen by fractional crystallization. However, the Basement Complex very largely consists of basic charnockitic granulites and amphibolites which represent pre-metamorphic intrusions of basaltic rock into a country rock that now consists mainly of intermediate charnockitic granulite and hornblende-biotite gneiss. Interbanded with these rocks is a series of paragneisses ; marbles, sillimanite- and kyanite-schists and quartzites. The field relationships suggest that the pre-metamorphic basic intrusives are related to the pyroxenite/perthosite complexes. Furthermore, it appears probable that the intermediate charnockitic granulites and hornblende-biotite gneisses represent a pre-metamorphic series of basaltic or andesitic volcanics and sediments, such as greywackes, derived from their erosion. If this contention should be proved correct, then the volumes of ultrabasic, basic, and perthosite rocks are not out of proportion to those that could be expected from the differentiation of basaltic or andesitic magma.

The remarkable similarity of the circular Mlindi and Chingali pyroxenite/perthosite complexes to the ring complexes of the Chilwa series, suggests that they have a common origin, even though separated in time by several hundred million years and a period of intense regional metamorphism. It is instructive, in this connection, to compare the Stormberg to Cretaceous igneous history of southern Nyasaland with the hypothesis suggested above for the derivation of the late Pre-Cambrian igneous rocks from basaltic magma.

In the Upper Palaeozoic period basaltic magma was available in depth over vast areas of Africa, as is shown by the huge swarms of dolerite sills and dykes and basalt flows of the Stormberg period. In Nyasaland the dyke swarms were particularly intense in the proto-rift and accounted for a crustal stretching exceeding one-fortieth in the

country between Blantyre and the Shire river, whilst sills of dolerite invaded the Karroo sediments of Chikwawa district. This stage of dyke and sill formation parallels the Pre-Cambrian vulcanicity which produced the basic charnockitic granulite and amphibolite intrusives of the Basement Complex.

Following the intrusion of the Karroo dolerites the huge masses of perthosite and nordmarkite that form the ring-intrusions of Zomba and Mlanje mountains were emplaced. With them came nepheline-syenites and carbonatite "ring complexes" of the Chilwa Series. Swarms of solvsbergite dykes were intruded at the time of the Chilwa vulcanicity. Pyrochlore and other rare-earth minerals were concentrated in these rocks. These stages are also represented in the Pre-Cambrian Basement Complex by perthosites, nepheline-syenites and ultrabasic types which lie in a similar setting. The Pre-Cambrian nepheline-syenites also contain pyrochlore and other rare-earth minerals. The coincidence could be hardly more remarkable and suggests that these Pre-Cambrian biotite-nepheline gneisses represent metamorphosed intrusions of nepheline-syenite and are not replacement rocks as has been suggested by Cooper (1946) and Bloomfield (1958).

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On Deformation at Thrust Planes in Shetland and the Jotunheim area of Norway.

By Derek Flinn

Abstract

In both north-east Shetland and the Jotunheim area of Norway two nappes of crystalline rocks occur lying one above the other and separated in time and space by orogenic sediments. These sediments were formed in part at least by erosion of the lower, and first emplaced, of the two nappes in both areas. Later emplacement of the upper nappes caused the deformation and metamorphism of these rocks. In both Norway and Shetland large quartzite conglomerates which occur immediately beneath the upper nappes have been deformed on patterns which have been determined from a study of the pebble shapes. From these patterns in both areas it is argued that the deformation was due to flow caused by pressure differences developed in the sole of the advancing nappe and possibly to extrusion from beneath the nappe.

In a recent account of the geology of north-east Shetland (Flinn, 1958), two nappes of meta-eruptive rocks were recognized. Comparison with the Norwegian geological literature seemed to indicate that the Shetland nappe pile is similar in many ways to the Jotun nappes in the Jotunheim area of Norway and this was confirmed by field investigations (Flinn, 1959).

In each area two relatively rigid nappes were produced from the same sources, and thrust one after the other to their present positions on thrust planes near the surface of the earth. In each case the lower of the two nappes was emplaced first and was subjected to rapid erosion due to the resultant relief. On the nappes sediments were deposited which had a similar history of formation and deformation but differed somewhat in materials because the source rocks in the two areas were different. The remainder of the first nappe and the newly deposited sediments were overrun by a second nappe in both areas and this caused the deformation and metamorphism of the rocks near the thrust plane.

In Shetland, the upper nappe rests on a schuppen zone which consists of tectonic slices of the newly deposited sediments (the Phyllite group), metagabbro and serpentine from the nappes, schists and gneiss from the metamorphic basement on which the nappe pile rests, and coarse hornblendic schists not found *in situ* in the area.

In Norway the upper nappe rests on the newly deposited sediments (the Valdres Sparagmite), on Cambro-Ordovician meta-sediments of geosynclinal facies forming the upper part of the lower nappe, and on plutonic rocks of the Bergen-Jotun kindred forming the lower part of the lower nappe.

A comparison of the metamorphisms associated with the emplacements of the nappes

In *North-east Shetland* a metamorphism (greenschist facies) is centred on the schuppen zones and confined to them and the rocks above and below them (Read 1934, Flinn, 1958). This metamorphism was constructive, and recrystallized and transformed most of the rocks in the schuppen zones. Sediments and meta-sediments, igneous rocks, and meta-igneous rocks have been transformed to phyllites with strong tectonite fabrics of similar symmetry and orientation (see later) indicating the same metamorphism for all. No mylonite or mylonite textures are found associated with this metamorphism and such cataclastic effects (occasional bending, straining, and fracturing of grains) as occur in these rocks can usually be shown to belong to a later destructive or dislocation metamorphism. The effects of a similar or possibly the same dislocation metamorphism are found in metamorphic rocks throughout Shetland.

In the *Jotunheim area* of Norway a metamorphism centred on the thrust at the base of the upper nappe has been described in some detail in the Bygdin and Bøverdal areas. At Bygdin (Strand, 1944, 1951b) the schistose base of the upper nappe grades upwards, sometimes through as much as several hundred metres, to gneiss and finally to unaltered massive plutonic rock. The rocks beneath the nappe, Valdres Sparagmite and Quartz-conglomerate, show the effects of a deformation decreasing in intensity downwards from the thrust planes. The rocks above and below the thrust plane have been frequently called mylonite, especially in older descriptions; and the metamorphism has been described as a mechanical one. In more recent descriptions, especially those by Strand, the very important role played by recrystallization in this metamorphism is emphasized. I have observed in hand specimens and more clearly in thin sections that while these rocks have some of the features of mylonites (fine banding and streaking of the constituents and eyes of relict minerals) the minerals other than those forming relict eyes have recrystallized as the rock was being mylonitized ; the rocks are blasto-mylonites with strong tectonite fabrics.

Similar rocks are found along the base of the upper Jotun nappe to the south of Lomskoll. Strand (1951a, p. 46 and 79) describes mylonite gneiss and phyllonites from south of Lomskoll formed by tectonization and recrystallization near the base of the nappe. Some of the specimens collected by me from the base of the nappe and below show very little cataclasis when examined in thin section ; minerals other than occasional large relics have recrystallized. Many of the specimens, however, are mylonites in which the fine-grained matrix has recrystallized but without increase in grain-size.

Some of the specimens I collected at the base of the upper Jotun nappe in Bøverdal are indistinguishable in thin-section from common regional metamorphic schists, while others are blasto-mylonites showing some mylonitic features. In these latter rocks large eyes of relict minerals lie in finely laminated, very fine-grained material composed of a micro-crystalline granular aggregate unlike the angular comminuted particles in true mylonites.

A comparison of the small structures produced near the bases of the nappes

In Shetland the pebbles in the Phyllite Group and the Funzie conglomerate below the upper nappe show a regional variation in shape (Flinn, 1956, 1958). In the Funzie conglomerate they vary from very flattened oblate spheroids (quartz-cakes) at the boundary thrust to tri-axial ellipsoids with the intermediate axis equal to the diameter of the equivalent sphere at some distance beneath the thrust, to tri-axial ellipsoids with the intermediate axis shortened some 10 per cent relative to the diameter of the equivalent sphere in the part of the conglomerate lying at the greatest exposed depth beneath the thrust. Other conglomerates beneath the upper nappe contain tri-axial ellipsoid pebbles with the intermediate axis shorter than the diameter of the equivalent sphere (Uriesetter, Fetlar and Ramnageo and Uyea, Unst).

The gradual change in shape of the pebbles in the Funzie conglomerate is accompanied by a change in the preferred orientation of the pebble axes, the shape and orientation of the quartz grains in the pebbles and in the matrix and of preferred lattice orientation of quartz and micaceous grains in the pebbles and in the matrix. The structure of the Funzie conglomerate was described earlier at some length (Flinn, 1956), and it was shown that these structures must have all developed at the same time as a result of the same metamorphism.

It was shown earlier (Flinn, 1958), that non-conglomeratic rocks and pebble-free bands in the conglomerates have fabric patterns similar in type, symmetry and orientation to those of nearby conglomerates. And it was argued that, although it is not possible to determine the deformation these non-conglomerate rocks have undergone as easily and exactly as for the pebbles in the conglomerates (by measuring them), it is possible to learn much about their deformation by comparison of their fabrics with those of nearby conglomerates. The same deformation must produce the same fabric symmetry whether or not there are pebbles present.

For instance, it is clear that the quartz-cake conglomerate at the thrust plane has been flattened ; it has been shortened parallel to the short axes (symmetry axes) of the pebbles and elongated equally in all directions at 90°. Rocks in the neighbourhood containing no pebbles

but having an otherwise similar fabric, i.e., a single schistosity, no lineation, coincident point concentrations of (001) of mica and [0001] of quartz, quartz grains flattened in the plane of the schistosity and so on (see Flinn, 1956) may be interpreted as also being the result of flattening. Extremely elongated pebbles have suffered constriction; they have been squeezed in the plane normal to their longest axes and as a result forced to flow in a direction parallel to their longest axes. The fabric features they have in common with non-conglomeratic rocks deformed in the same way are coincident girdle concentrations of (001) of mica and [0001] of quartz lying in the plane of compression, elongate grains orientated parallel to the girdle axis (lineation) and signs of flow parallel to that axis. More common are conglomerates with pebbles of tri-axial allipsoid shape and fabrics intermediate to those described above. Non-conglomeratic rocks of similar fabric are common.

Those rocks which have suffered some degree of constriction, that is those rocks whose pebbles, if they contained pebbles, would show a shortening of the intermediate axis relative to the diameter of a sphere of the same volume, and which are layered have been folded and the fold axes are invariably parallel to the elongation direction and girdle axis (Flinn, 1958). The folds vary in wavelength from millimetres to decametres and although individual folds are often monoclinic due to overturning, the folding as a whole is orthorhombic due to overturning in opposite directions.

A study of the rocks in north-east Shetland in the light of the discussion given above shows that the deformation below the nappes is in general the result of constriction (Flinn, 1958). Only in the south-east corner of the Funzie conglomerate was deformation due to flattening. The relatively high symmetry of both the constricted and the flattened rocks (orthorhombic to axial) is an indication that the forces causing the deformation acted parallel to the symmetry axes of the rock fabric. The greatest, intermediate and least stresses were parallel to the least, intermediate and longest axes of the pebbles respectively.

In *Norway* deformed conglomerates have been described from a number of localities in the Jotunheim area. It is necessary to repeat these descriptions to some extent in order to emphasize features which the original authors did not emphasize and in order to introduce in the right context some new observations of my own. The best described occurrence is the meta-conglomerate at Bygdin (Strand, 1944). There the strike of the base of the upper nappe changes abruptly from south of east to almost south, while at the same time the dip changes from steep to the north to 35° to the west (Strand, 1944, fig. 1, p. 15). At the hinge there is a narrow invagination of the underlying quartz-conglomerate (Valdres Sparagmite) into the nappe. In the invagination the

pebbles in the conglomerate show extreme elongation and nearly equidimensional cross-section ; dimensions of 80:1.5:1 have been given by Goldschmidt (1916). A recently formed exposure in the tip of the invagination showing the contact between Jotun rocks and the conglomerate contains pebbles with equidimensional cross-sections elongated to rods of great length. Along the thrust to the south of the invagination Strand reports the occurrence of pebbles of quartz-cake type. The same type of pebble occurs along the thrust to the east of the invagination. To the south-east of the invagination and away from the thrust, the pebbles are usually rather elongate tri-axial ellipsoids. Strand (1944, p. 19) gives some pebble dimensions from here, and most of the intermediate axes are smaller than the diameter of an equivalent sphere.

I found the shore line of the lake to the east of Bygdin to provide an instructive section through the conglomerate. In the north at the contact with the Jotun rocks the pebbles are of quartz-cake type. Towards the south they become progressively elongated at the expense of the intermediate axis so that at the southern margin of the conglomerate on the shore line the pebbles are slightly flattened prolate ellipsoids. The elongation is not nearly so great as in the invagination ; deformation decreases away from the bottom of the nappe. This change in shape of the pebbles and decrease in intensity of deformation of the pebbles away from the thrust is similar to the change in shape of the pebbles and decrease in intensity of deformation away from the thrust observed in the Funzie conglomerate.

In Norway the fabric of the matrix of the conglomerate reflects changes in shape of the pebbles in the same way as it does in Shetland. Strand (1951, p. 30) notes that where there is no distinction between the intermediate and least axes of the pebbles there is no schistosity. The Jotun rocks at the contact with the conglomerate have similar fabric patterns to the adjacent conglomerates. Near the invagination, where the pebbles form long rods with equidimensional cross-sections, I found that the Jotun blasto-mylonites frequently contain spindles of feldspar with equidimensional cross-sections elongated in the same direction as the pebbles. Like the adjacent conglomerate, such rocks have no schistosity only a lineation parallel to the elongation direction (a girdle fabric). Along the contact to the east and south of the invagination, where the conglomerates are schistose due to the flattening of the pebbles, the adjacent Jotun blasto-mylonites are also schistose but have a tendency to develop a lineation by streaking out of the constituents.

To the east, south-east, and south of Bygdin, Strand reported folds which were usually on a small scale in the rocks below the nappe. The fold axes generally trend either north-east or north-west. The

north-east (Caledonoid) folds are mainly isoclinal and overturned to the south-east, although in some places they are overturned to the north-west. The north-west (cross) folds are generally symmetrical but he observed overturning which was more often to the south-west than to the north-east. The axes of the cross-folds are parallel to the elongation direction of the conglomerates, and in some places the conglomerates and even the individual pebbles are folded on these axes.

In Bøverdal Landmark (1948) described how the structure in the blasto-mylonites in the lower part of the nappe changes upwards from a schistosity parallel to the base to either a lineation trending and plunging south-eastwards, or a steep schistosity with a south-easterly strike. The lineation according to Landmark is due to a linear orientation of mineral grains and aggregates, or a " stretching " direction without megascopically visible orientation of the mineral grains. Light and dark streaks in the rocks appear as eyes in sections perpendicular to the lineation. The two schistosities in the base of the nappe occur in the Valdres Sparagmites below the nappe, but again not together. A very prominent " stretching "-type lineation also occurs, parallel to the lineation in the overlying nappe and also to a lineation tin the upper part of the basement to the west. By means of field investigation and examination of orientated thin-sections and polished surfaces of rocks collected from the area I confirmed that the lineated rocks described by Landmark have the same mica girdle fabric as the lineated rocks at Bygdin. A feature of the area is the occurrence everywhere of tension fractures about normal to this lineation. A quartz-conglomerate occurs some distance below the nappe with its pebbles elongated parallel to the lineation and somewhat compressed parallel to the intermediate axis (15:3:1). As in Bygdin, folds on both north-east and north-west axes occur. Those on north-westerly axes (parallel to the lineation) are not overturned consistently in any one direction throughout the area.

Along the base of the upper Jotun nappe to the south of Lomskoll the lineation appears from observations I made during a very short visit to the area to be less constant in direction than in the other two areas. The thrusts and the layering in the rocks dip 45° to the south or south-west, while the lineation at some structural levels is parallel to the strike, and at others more nearly perpendicular to the same strike. The lineation in both these directions, like that in the rocks of the other two areas, is due to the parallel orientation of elongate minerals, girdle and partial girdle fabrics of micaceous minerals, and the streaking out of grain groups: it is found in metamorphic derivatives of both the crystalline rocks of the nappes and the metasediments below the nappes.

The lineations in the rocks near the base of the thrust in Bøverdal

and to the south of Lomskoll are due to the same fabric patterns as the lineations in the rocks at Bygdin. At Bygdin the shapes of the deformed pebbles indicate the pattern of the deformation that gave rise to the lineation. In Bøverdal and to the south of Lomskall pebbles are absent, or almost absent from the rocks, but comparison of the fabrics with those of the Bygdin area in the manner described above for Shetland shows that in all three areas the deformation was of similar type. Thus, it seems probable that where the rocks are schistose and not folded (quartz-cake pebbles in conglomerates) flattening has taken place ; and where the rocks have no schistosity but are lineated owing to a girdle arrangement of the poles of (001) cleavage of micaceous minerals, constriction has taken place. Thus at Bygdin, in Bøverdal and to the south of Lomskoll the rocks above and below the base of the nappe have suffered flattening in some places and constriction in others. Flattening is mostly found at the boundary between the nappe and the underlying rocks. Constriction is much more common and is found both at the boundary and away from it.

The fabrics in the conglomerate and the non-conglomeratic rocks at the base of the upper Jotun nappe are so similar to those at the base of the Shetland nappes that it must be concluded that they are not only due to similar patterns of deformation, but that the causes of those deformations must be similar.

A comparison of the tectonics of North-east Shetland and the Jotunheim area

In both areas the deformations and the metamorphisms that accompanied them seem from their close geographical association with the thrusts to be genetically related to the thrusting. But from the nature of the deformation patterns described above it is clear that the relationship cannot be as simple as has been suggested in the past. Oftedahl (1948) suggested that the principal component in the deformation was simple shear movement of thrusting, evidently extending upwards and downwards from the thrust plane. If this hypothesis were correct the deformation above and below the thrust plane should be of simple shear type and should decrease upwards and downwards from it. From observations of pebble shapes in both Shetland and Norway it is clear that the deformation in the rocks in contact with the thrust plane is either flattening giving rise to disc-like pebbles, or constriction giving rod-like pebbles, instead of simple shear giving tri-axial ellipsoid pebbles with their intermediate axes equal to the diameter of the equivalent sphere (biaxial strain). At greater depth below the thrust in both areas the pebbles are tri-axial ellipsoids but not in general of bi-axial strain type. In Shetland it was shown that the deformation in the Funzie conglomerate increases from the thrust

plane to a maximum at some distance below the thrust plane before it begins to decrease. Oftedahl considered the deformation in terms of a simple shear component due to thrusting and another component of unspecified type and origin. It is clear from the way the intensity of deformation changes with depth that there must be a component due to simple shear, but it is too small for its effects to be recognizable in the fabric of the rocks.

Kvale (1945, 1953) working in an area to the south-west of the Jotunheim mountains, which is somewhat similar tectonically, considered fabrics similar to some of those described above (constriction fabrics) to be due to stretching. He considered that stretching can be partially explained by the rocks being dragged out by faster moving overlying nappe, an explanation that seems to be essentially similar to that of Oftedahl.

Balk (1952 and 1953) has likened the deformation of the rocks beneath the advancing nappe to the deformation of metal in a rolling mill. However, cold rolled metal elongates normal to the roller axes and shows no change of dimension parallel to the roller axes. Substances like putty and plasticene show some elongation parallel to the roller axes as well as normal to them but never shortening as may be demonstrated with a mangle.

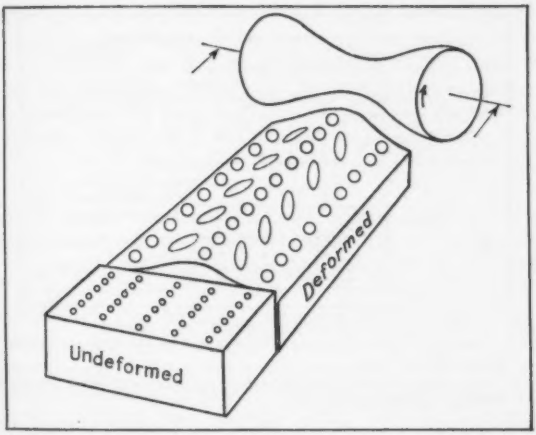
Balk observed folds whose axes he believed lay in the movement direction, that is normal to his hypothetical roller axes. He suggested that the shortening parallel to the roller axes causing this folding was due to inhomogeneities in the rocks (Balk, 1953, 99). This explanation is insufficient ; in the Bygdin and Funzie conglomerates both elongation and shortening parallel to the " roller axes " occur in close proximity despite the great homogeneity of these rock masses.

Experiments were carried out to test whether rolling of putty-plasticene conglomerates by rollers with an annular groove might produce elongation normal to the roller axes in the ridges formed on the conglomerate and flattening in the troughs (and thus explain the relations at Bygdin). In fact, maximum elongation was observed on the sides of the ridge produced by rolling, with a direction obliquely foreward from the bottom of the trough towards the tops of the ridges (Text-fig. 1). This experiment appears to have no analogy with the rock deformation under consideration, since such patterns have not been found in the field.

In further experiments deeply grooved glass plates were slid parallel to the groove over syrup-air and vaseline-treacle model conglomerates, the air and the treacle being dispersed in the form of small spheres through the syrup and vaseline respectively. In these experiments the shear plane between the glass plate and the model conglomerate continued across the entrance to the groove so that the material in the

groove remained undeformed and was carried bodily foreward with the glass plate (Text-fig. 2a). At Bygdin the quartz-conglomerate in the invagination suffered intense constriction.

All the hypotheses, analogies, and experiments mentioned above assume that the deformation of the rocks resulted from a continuation into the rocks of the monoclinic shearing movement on the thrust plane itself. It is clear from the field association of the deformation with the thrust planes in Shetland and Jotunheim that the metamorphism is genetically related to the thrusting, but this does not mean that the

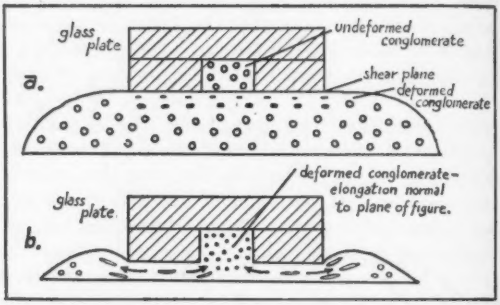


TEXT-FIG. 1.—Deformation of a conglomerate by a roller with an annular groove.

deformation is necessarily a part of the actual thrusting. The close association at Bygdin of both flattening and intense constriction immediately beneath the nappe, and the general lack of monoclinic symmetry with suitably orientated symmetry plane in the tectonite fabrics are evidence that the rocks near the thrust were deformed by patterns of movements differing very considerably from those in the actual thrust plane. The initial increase in intensity of deformation downwards from the thrust plane in Shetland shows that here at least the deformation below the thrust was not caused by movement on the immediately overlying thrust plane.

The deformation in the rocks below (and in Norway also above) the thrust must be considered independently of the thrusting, in order to determine its true origin. This deformation may be caused only indirectly by the thrusting, instead of directly as has been generally assumed hitherto. Movement of a nappe on an uneven thrust plane or

slide would, in places, cause depressions in the base of the nappe to coincide with elevations on the surface of the underlying rocks and vice versa, thus giving rise to areas of high and low pressure in the thrust plane. Flow from high pressure areas to low pressure areas could give rise to flattening in the high pressure areas. In rocks farther removed from the thrust plane, longer-range pressure gradients might cause flow of a more cylindrical nature connected, in the case of nappes travelling over the surface, with extrusion at the front and even the sides of the nappe. Such flow could take place plastically during the



TEXT-FIG. 2.—Deformation of a conglomerate by a nappe
(*a*) glass plate (nappe) sliding over conglomerate parallel to a groove in its base.
(*b*) glass plate (nappe) with groove in its base being pressed down without lateral movement.

advance of the nappe, or more slowly as creep (rheid flow, Carey, 1953) after the emplacement of the nappe or during pauses in its emplacement. Flattening and constriction fabrics and fabrics transitional to these would be expected to form under these circumstances, but not as a result of simple shearing (thrusting). Such fabrics are observed in Shetland and Jotunheim, while fabrics indicative of simple shearing are not.

In an attempt to explain in the light of the ideas presented above the intense constriction in the invagination at Bygdin, deeply grooved glass plates were pressed down without lateral movement on syrup-air and vaseline-treacle conglomerates. Flow beneath the glass plate took place, but did not occur at all in the groove until the depth of the model conglomerates beneath the base of the glass plate had been reduced to less than the depth of the groove (Text-fig. 2*b*). Then strong flow and elongation commenced in the groove parallel to its length and simultaneously occurred beneath the plate in a direction normal to the groove. The depth of the deformed conglomerate beneath

the upper nappe at Bygdin is considerably greater than the depth of the invagination and there is no sign of elongation in a direction normal to the elongation in the invagination in the rocks on either side. The intense elongation in the invagination at Bygdin is probably not due to channelled flow.

This conclusion is supported by the fact that the junction between the Jotun rocks and the underlying conglomerate, well exposed in stream and road sections near the Bygdin Hotel, marks a change in lithology and not a structural break ; there is no change in structure across the junction and no structural break at the junction. Although this junction must at one time have been the thrust plane it is no longer one. It is possible that the nappe advanced over an increasingly thick layer of its own mylonite material. Once formed this material would behave like the sediments below the nappe ; it would not transmit the simple shear deformation of thrusting but, like the sediments beneath it, it would tend to flow towards low pressure areas, and fabrics similar to those formed in the sediments would be formed. The fabrics of the specimens I collected from the " mylonite " zone at the base of the nappe were similar to those in the rocks below the nappe. The blasto-mylonite around the invagination at Bygdin, like the conglomerate in the invagination, has suffered intense constriction. This constriction must have taken place at a time when the thrust plane had already risen to some height above the base of the nappe by conversion of the rocks to blasto-mylonite. If as a result of the constriction the rocks flowed in a direction slightly oblique to the conglomerate-" mylonite " boundary that surface would be " folded ". Thus the invagination may be the result of constriction and not vice versa. An investigation of the fabrics in the base of the upper Jotun nappe is needed to settle these and other problems.

It has long been clear that for nappes of the size considered here to have moved into their present positions, the friction opposing their movement must have been very low (Oftedahl 1943, Kvale 1945, Hubbert and Ruby 1959, p. 122). Hubbert and Ruby have explained how frictional drag on the base of the nappe would be greatly reduced if fluid pore pressure approaching the pressure due to the weight of the overlying nappe were present in the rocks cut by the thrust plane. It is possible that such a state prevailed beneath both the Shetland and Jotun nappes, since in both areas nappes of relatively impervious crystalline rocks were thrust onto newly deposited sediments. The high pore pressure produced by the passage of an impervious nappe over unconsolidated sediments at a rate faster than the water could escape would not only reduce the friction on the thrust plane, but also the strength of the sediments (and blasto-mylonites) making it easier for them to deform by flowing to areas of lower pressure.

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The Occurrence of Hypersthene in Hawaiian Basalts

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ABSTRACT

The occurrence of hypersthene in Hawaiian basalts, related broadly to the lime-silica values in the bulk analyses, is now considered and graphically expressed in terms of the quantitative proportions of the normative constituents, diopside, hypersthene, olivine, and quartz in analyses of Kilauea and Mauna Loa lavas.

IN AN earlier paper (Tilley and Scoon, 1961) the occurrence of hypersthene in Mauna Loa as distinct from Kilauea basalts was shown to be related broadly to the lime-silica values in the bulk analyses of these rocks, and it was noted that the few Kilauean lavas that carried hypersthene were the more siliceous types, lower in lime (Tilley and Scoon, 1961, p. 62).

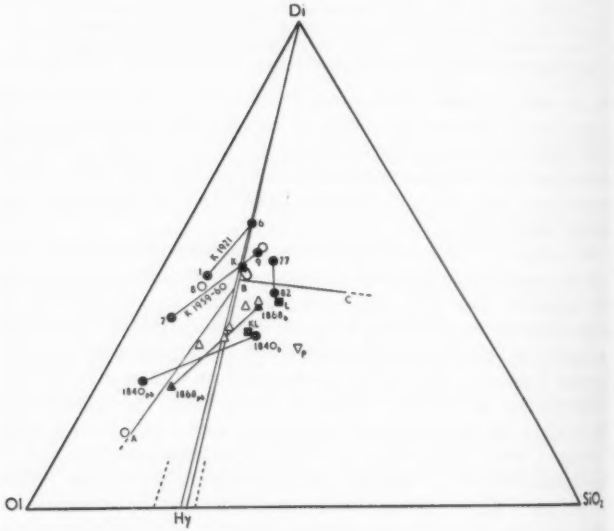
The problem of the occurrence of hypersthene in these basalts can now be considered further with a procedure which attempts to evaluate the analytical data by concentrating attention on the quantitative proportions of the normative minerals diopside, hypersthene, olivine or quartz, the lime of normative anorthite thus being excluded. For this purpose normative hypersthene is expressed as such or alternatively converted to equivalent olivine + quartz and the constituents Di + Ol + Qz are computed to 100 and shown in a ternary plot. Text-fig. 1 presents the data on a suite of Kilauea and Mauna Loa basalts and in Text-fig. 2 the two three-phase assemblages (Opx + Cpx + Ol, Opx + Cpx + Qz) are ideally presented with the thickened lines along the join Di–Hy portraying the extent of solid solution under conditions of a hypersthene-clinopyroxene paragenesis in a crystallizing basaltic magma.

Text-fig. 1 presents the actual normative data with the lines AB, BC, giving the empirically derived boundary limiting the field of hypersthene (below ABC). The departure of these lines from the joins to Ol and Qz as shown in Text-fig. 2 is to be ascribed, doubtless, to procedural methods adopted in the ordinary norm calculation. Text-fig. 3 shows the results in the alternative form where hypersthene in the norm is preserved as such in the calculations. The equations of the lines AB and BC are indicated in the diagram. As an illustration of the procedure of plotting we may set out the analysis of the 1887 Mauna Loa hypersthene basalt and its clinopyroxene (Table 1).

In this example the normative composition of the clinopyroxene itself plots close to the limit B on the join Di–Hy of Text-figs. 1 and 3 approximately 53 weight per cent hypersthene. In contrast, one of the analysed clinopyroxene fractions of a 1921 Kilauea lava without hypersthene has a corresponding plot at Di 60·4, Hy 27·3, Ol 12·3

(or Di 60·4, Ol 32·1, Qz 7·5) thus like the plots of the 1921 lavas themselves (Text-fig. 1) remote from the limiting region at B (Tilley, 1960, Table 1, p. 50).

The results expressed in the foregoing account need to be tested



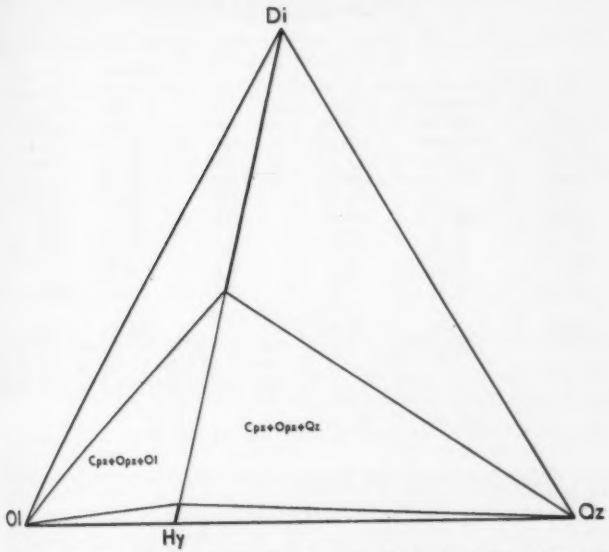
TEXT-FIG. 1.—Plot of normative compositions (Di + Ol + Qz = 100) of Kilauea and Mauna Loa basalts showing the field of rocks with modal hypersthene (below the line ABC)

Di = diopside, Ol = olivine, Hy = hypersthene (normal range, the extremes of En and Fs being indicated by the ends of the dotted lines right and left of Hy).

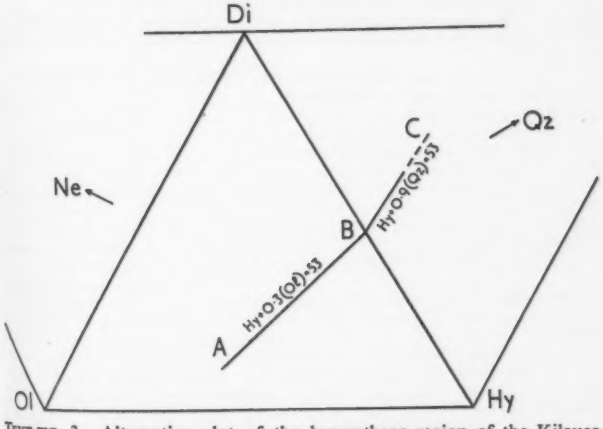
Circles = Kilauea lavas, Triangles = Mauna Loa lavas.

Filled squares : averages for Kilauea (K), for Mauna Loa (L) and Koolau basalts, Oahu (KL). P = Palolo dyke (Oahu). The joins 1840_{pb}–1840_{b}, K 1921, and K 1959–60 represent Kilauean eruptions of the year indicated. The join 77–82 represents Kilauean eruptions of 1955. The join 1868_{pb}–1869_{b} refers to Mauna Loa lavas of the 1868 eruption.

further with the accumulation of additional first class analyses of Hawaiian tholeiitic basalts as they come in. It is not to be expected that the results themselves can be applied directly to the analytical data of tholeiitic lavas of other volcanic fields.



TEXT-FIG. 2.—Idealized diagram of the system Di-Ol-Qz showing the two three phase fields (Opx-Cpx-Ol), (Opx-Cpx-Qz).



TEXT-FIG. 3.—Alternative plot of the hypersthene region of the Kilauea-Mauna Loa lavas in terms of normative Di-Ol-Hy and Di-Hy-Qz. The point B is at 53 weight per cent (Hy).
Ne = nepheline.

TABLE I

	1	2	Norms	1	2
SiO_2	51·77	53·00			
Al_2O_3	13·78	5·30	Qz	1·98	1·02
Fe_2O_3	2·04	1·14	Or	2·22	1·67
FeO	9·13	11·21	Ab	19·91	7·86
MnO	0·16	0·22	An	25·85	9·45
MgO	8·58	15·65	Di	18·97	35·68
CaO	10·12	10·81	Hy	24·58	40·34
Na_2O	2·36	0·96	Il	3·34	1·82
K_2O	0·39	0·25	Mt	3·02	1·62
H_2O+	0·07	n.d.	Ap	0·34	—
H_2O-	0·06	0·12	H_2O+	0·13	0·12
				100·34	99·58
TiO_2	1·76	0·95			
P_2O_5	0·17	—	Di 41·7 Di 41·7	Di 46·3 Di 46·3	
			Hy 54·0 Ol 39·2	Hy 52·4 Ol 37·9	
			Qz 4·3 Qz 19·1	Qz 1·3 Qz 15·8	
	100·39	99·61			

1. Hypersthene basalt (1887 flow of Mauna Loa), Tilley and Scoon, 1961, Table 3, No. 4.
2. Clinopyroxene of 1, analyst, J. H. Scoon.

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DEPARTMENT OF MINERALOGY AND PETROLOGY,
CAMBRIDGE.

CORRESPONDENCE

ANTIQUATONIA MOLARUM NOM. NOV. PRO *PRODUCTUS ROTUNDUS* GARWOOD

SIR,—Muir-Wood and Cooper (1960, p. 271) having shown that *Productus rotundus* Garwood, 1913 (p. 569, pl. li, figs. 3a, b), a species referred by them to the genus *Antiquatonia* Miloradovich 1945, is preoccupied by *Productus rotundus* Pander, 1830 (p. 85, pl. xxi, figs. 6a–d), the name *Antiquatonia molarum* is here substituted for the former (*molae*, feminine plural, a mill, in reference to Fawcett Mill, Shap District, Westmorland, whence Garwood described his species).

J. SELWYN TURNER.

DEPARTMENT OF GEOLOGY,
THE UNIVERSITY,
LEEDS, 2.
17th February, 1961.

REFERENCES

GARWOOD, E. J., 1913. The Lower Carboniferous succession in the north-west of England. *Quart. Journ. Geol. Soc.*, lxviii, 449–582.

MUIR-WOOD, H., and COOPER, G. A., 1960. Morphology, classification and life habits of the Productoidea (Brachiopoda). *Geol. Soc. Amer., Mem.* no. 81.

PANDER, C. H., 1830. *Beiträge zur Geognosie des russischen Reiches.* St. Petersburg.

INHALANT AND EXHALANT FEEDING CURRENT SYSTEMS IN RECENT BRACHIOPODS

SIR,—In his recent article on feeding mechanisms of spire bearing brachiopods Dr. M. J. S. Rudwick (1960) divided spirolophes into two groups, one possessing an inhalant filter feeding current-system and the other characterized by an exhalant system. He noted (pp. 380, 381) that the exhalant type is unknown in living brachiopods. In subsequent correspondence Professor Alwyn Williams (1960, p. 516) considered that the absence of an exhalant system in any living species may reflect a fundamental disadvantage of the system and reiterated his views on the deuterolophous nature of the atrypaceid and spiriferoid lophophore.

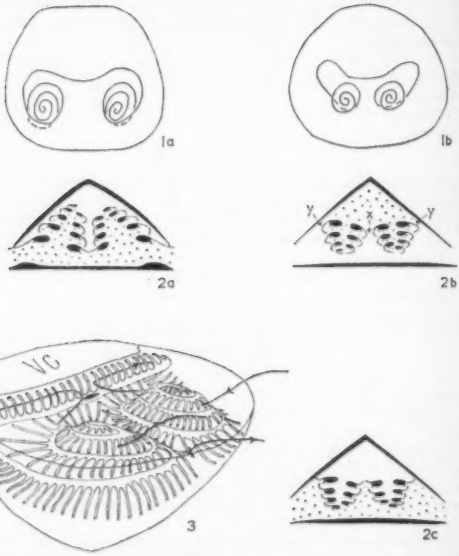
It is not the purpose of the present communication to enter into the discussion on the nature of the lophophores of fossil spire bearers or into the vexed question of the function of the jugum, but to consider the probable nature of the current system of the recent genus *Discinisca* and its implications.

As Rudwick and Williams rightly point out an exhalant system has not been noted in any recent genus and from what is known of the morphology of other recent articulate genera which have not been examined experimentally, such a system is unlikely to occur. This does not appear to be the case in the inarticulate genus *Discinisca*, which is represented in the sea to-day by several species. The gross morphology of the lophophore of this genus is known in some detail through Blochmann's studies (1900), but the lophophore has not been examined functionally.

From Blochmann's figures and description it is seen that the lophophore of *Discinisca* is spirolophous but differs from that of *Crania* in having the apices of its cones ventrally directed (cf. figs. 1a, 1b). As in the case of *Atrypa* and *Spirifer* described by Rudwick this difference is more fundamental than it might appear initially. Looking up the left spiralium from its base towards the apex, the lophophore is coiled anti-clockwise in *Crania* and clockwise in *Discinisca*. Thus *Discinisca*, in terms of its direction of coiling of the lophophore, belongs to Rudwick's " *Spirifer* group."

Rudwick (p. 373) has shown that the filaments of the left brachium, when viewed along the brachial axis towards the tip of the brachium, filter in a clockwise direction and those on the right brachium in an anti-clockwise

direction. The water always flows from the brachial lip across the groove to the filaments. He has stated (p. 376) that for an efficient filter feeding system the filtered and unfiltered water must be kept separate and that distinct exhalant and inhalant chambers must be formed. He has then assumed that fossil spire bearers had such an efficient system, an assumption which appears justifiable in view of what is known of the feeding mechanisms of recent brachiopods.



TEXT-FIG. 1.—Diagram of dorsal valve of (*a*) *Crania anomala* to show dorsally directed lophophore spires. (*b*) *Discinisca lamellosa* to show ventrally directed lophophore spires.

TEXT-FIG. 2.—Transverse sections to show filter feeding systems (*a*) *Crania anomala* (Rudwick, 1960), (*b*) *Discinisca lamellosa* reconstructed with inhalant system (*c*) *Discinisca lamellosa* reconstructed with exhalant system. Unfiltered water stippled.

TEXT-FIG. 3.—Diagrammatic view into dorsal valve of *Discinisca lamellosa* to show disposition of filaments when reconstructed with an exhalant filter feeding system. Only one row of filaments shown. Arrows indicate direction of currents. VC-visceral cavity.

Applying Rudwick's assumption to the lophophore of the recent *Discinisca* it is seen that two possible reconstructions can be made. The filter feeding current-system could be inhalant (Fig. 2b) with the filaments of the proximal whorl extending onto those of the whorl beneath, or it could be exhalant (Fig. 2c) with the filaments of the proximal whorl extending to the dorsal mantle and the space between the proximal whorl and the whorl beneath closed by filaments from the latter whorl. Reconstruction 2b does not produce an efficient inhalant system since the separation of inhalant and exhalant chambers is incomplete at X and Y. In *Crania* (Fig. 2a), these spaces are closed by

filaments as the lophophore is coiled into dorsally instead of ventrally directed spirals. Thus on the grounds of functional necessity for an efficient filter feeding current-system the recent inarticulate *Discinisca* should possess an exhalant system. The filaments in such a system could be arranged as in fig. 3. The filaments of the proximal part of the lophophore flexed ventrally to touch the anterior body wall, forming a tunnel to conduct that part of the exhalant water, which has passed through these filaments to the main exhalant chamber in the spirals.

Cases when scientific method can be applied rigorously in palaeontology are perhaps rather rare, but this would appear to be one such case. The theoretical conclusion that *Discinisca* possesses an exhalant system is capable of being verified by direct observation on the living animal when material becomes available. If it does possess an exhalant system it will strengthen the value of Rudwick's hypothesis for fossil spire bearers with spirolophous lophophores. It will not of course prove that fossil spire bearers had spirolophes rather than deuterolophes.

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RUDWICK, M. J. S., 1960. The feeding Mechanisms of Spire-bearing Fossil Brachiopods. *Geol. Mag.*, xcvii, 369–383.

WILLIAMS, A., 1960. Feeding Mechanisms of Spire-bearing Brachiopods. *Geol. Mag.* xcvii, 514–516.

A. J. ROWELL.

DEPARTMENT OF GEOLOGY,
UNIVERSITY OF NOTTINGHAM.
18th April, 1961.

ORIGIN OF ALBITE PORPHYROBLASTS

SIR,—In his paper on the " Origin of Albite Porphyroblasts in Rocks of the Ben More–Am-Binnein Area ", in the January–February issue (*Geol. Mag.*, 98, 41–55), Dr. Jones has misrepresented, to an extent worth correction, the conclusions which I drew from work on albite gneisses in this area and elsewhere in the Dalradian some time ago (Trendall, 1953).

First, a minor point on page 43: Dr. Jones is confusing my views with Reynolds' in the sentence " The origin of the porphyroblasts was considered by Trendall to be due to a soda metasomatism, possibly from a trondhjemitic magmatic source." I thought the most likely source of soda to be a migmatite, similar to those of Belmullet and Foxford in Ireland and Cromar and Glen Shee in Scotland, underlying the crestal area of the Cowal anticline.

On page 51 I am quoted as an authority for the assertion that the albite schists are confined " to the Ben Ledi Grit Group, both in the present area and in the rest of the South West Highlands and Co. Mayo ". On the contrary, I wrote: " In the Dalradian of North Mayo albite schists occur throughout a great thickness of Dalradian rocks whose correlation with the Standard Perthshire Succession is uncertain, but which are likely to lie below the Loch Tay Limestone, and possibly extend downwards as far as the Schiechallion Boulder Bed. There is no suggestion that in Mayo the albite schists are confined either to (i) Bed Ledi Grits or their equivalents, or (ii) rocks which may have contained abundant felspathic grits."

Dr. Jones fails to make clear where his views differ from mine, and does not in fact suggest any origin for the albite porphyroblasts, except to agree with Reynolds and myself that " sodium was introduced from an extraneous source by a large scale metasomatism ". He proposes, it is true, that " the formation of albite schists is an integral part of the regional metamorphism " but suggests that most of the soda " arose as the result of a late regional

soda metasomatism ". This appears to be close to my own view, quoted by Dr. Jones, of a " late para-tectonic phase of metamorphism to some extent unconnected with the earlier regional metamorphism ", which I followed with a proviso " that some influence on the metasomatic metamorphism is exerted by the regional metamorphism, which suggests that there is no great difference between the two phases ".

My main reason for separating the late soda metasomatism from the main regional (zonal) metamorphism was the discordance between the albite and garnet " isograds ". Dr. Jones apparently does not accept, as I did, the inversion of the zonal metamorphism in this area. The garnet isograd on his map (Text-fig. 1) indicates a near-vertical surface, since it crosses several deep valleys without deviation; the limit of albite porphyroblasts, on the other hand, is a gently curved surface following the predominant regional foliation—the Cowal anticline. For two manifestations of the same regional metamorphism to possess bounding surfaces intersecting at a high angle a considerable time gap seems to be necessary.

Perhaps it is mainly a question of nomenclature: how late must a soda metasomatism be before it ceases to be considered an integral part of the regional metamorphism? Further details will no doubt appear in the later publication to which Dr. Jones refers, including a suggestion, it is to be hoped, concerning the source of the soda; meanwhile there seems to be no significant difference between Dr. Jones' views and my own on the problems involved.

A. F. Trendall.

Geological Survey Department,
P.O. Box 9,
Entebbe, Uganda.
25th April, 1961.

REFERENCE

Trendall, A. F., 1953. The Origin of Albite Gneisses. *Ph.D. Thesis, University of Liverpool.*

REVIEW

Edinburgh Geology, An Excursion Guide. Edited by G. H. Mitchell, E. K. Walton, Douglas Grant. xv + 222 pp., 8 plates and numerous figures. Oliver and Boyd, Edinburgh and London. Price 12*s.* 6*d.*

This guide compiled and edited by Fellows of the Edinburgh Geological Society, describes twenty whole or half day excursions which can be made from Edinburgh. It includes much detailed local information about the Arenig to Upper Carboniferous stratigraphy and the wide range of igneous and glacial phenomena of the east and central parts of the Midland Valley and Southern Uplands. Understandably in a book of this scope, the only general attempt to synthesise the information is a brief introductory chapter. Some of the chapters include much unpublished material and all contain lists of references. In planning and carrying out field work, this little book will be invaluable not only to experienced geologists but also to those with less experience, for whom a glossary of terms and a table of mineral compositions are included. The maps, photographs, and general format are most attractive and the price seems remarkably low.

P. F. F.

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